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The Geological Society of America Bulletin

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NEW CODE OF STRATIGRAPHIC NOMENCLATURE

Authors of papers to be submitted to GSA for publication are requested to follow the usage for stratigraphic nomenclature recommended in the new Code of Stratigraphic Nomenclature (See May 1961 issue of the Bulletin of the American Association of Petroleum Geologists).

R. H. DOTT, JR. *University of Wisconsin, Madison, Wis.*

Squantum "Tillite", Massachusetts—Evidence of Glaciation or Subaqueous Mass Movements?

Abstract: The writer interprets the Boston Bay group as partially intertongued lithofacies of coarse conglomerate and associated volcanic rocks (Roxbury conglomerate) and finely laminated siltstone and mudstone (Cambridge "slate"). He regards the Squantum "tillite" as multiple discontinuous lenses of unsorted pebbly to bouldery mudstone of various ages interstratified within this complex. Although fossil evidence is inconclusive, stratigraphic relationships and clastic compositions suggest that these deposits were formed during middle Paleozoic diastrophism. The writer does not consider the Squantum as a temporal equivalent to southern hemisphere Permo-Carboniferous glaciation, as advocates of a glacial origin had supposed. Lack of glacial pavement and comparison of composition, graded bedding, gross stratigraphy, and tectonic setting with known submarine slide and mass-flow deposits render the glacial interpretation suspect. Origin by gravity movement of rapidly deposited, volcanic-rich sediments and periodic resedimentation by turbidity currents in a Paleozoic eugeosynclinal belt is a more plausible alternative.

Very poor sorting, striated clasts, faceting, and even some rafted erratic fragments in fine, laminated mudstones can arise from mechanisms other than glaciation. A preserved, extensive, grooved and polished pavement overlain by poorly sorted, till-like material—particularly if nonmarine—is the most compelling glacial evidence. Very large erratic boulders are suggestive of ice movement, as are *abundant* rafted erratic fragments in fine muds, as in the Gowganda formation of Ontario. Independent biologic or isotopic cold-temperature indicators are sorely needed to strengthen glacial interpretations. General stratigraphic relationships and tectonic setting serve as important factors in judging probability of alternate interpretations, particularly in geosynclinal sequences. From these criteria, the writer judges that only the Permo-Carboniferous glaciation in certain parts of the southern hemisphere is firmly established. The Gowganda and some other Precambrian deposits are very likely glacial, but most postulated examples must be re-evaluated.

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INTRODUCTION

Sutherland (1870) recognized what he believed to be ancient Carboniferous glacial deposits in South Africa (the Dwyka tillite); they have great areal extent and rest upon a "pavement" of polished and grooved older rocks. The Dwyka has become the best documented of all supposed pre-Pleistocene tillites. Present position near the equator of the Dwyka and presumably similar deposits on other southern continents has supported the hypotheses of continental drift and polar wandering.

After 1870, increasing numbers of boulder beds all over the world were interpreted as ancient glacial tillites (*See* Coleman, 1926). In 1875 W. W. Dodge (*in* Mansfield, 1906) proposed a glacial origin for the Squantum rocks near Boston, Massachusetts which, with the Precambrian Gowganda formation of western Ontario (Coleman, 1907) and the Precambrian or Cambrian "tillite" near Salt Lake City, Utah (Blackwelder, 1932), constitute North America's most famous supposed ancient tillites. The Squantum was thought to be equivalent to the Dwyka (Sayles and LaForge, 1910), but it is unusual in its isolated geographic position, as is a possible late Paleozoic tillite in the northern Ural Mountains (Backland, 1929). These are particularly troublesome to advocates of continental drift, because plotting of their predrift positions brings both rather close to the equator. Past positions of North America and Europe closer to the equator are, incidentally, more in accord with other Paleozoic evidence, such as paleomagnetic data, extensive coals, and somewhat younger evaporite deposits, than are their present positions.

This paper suggests an alternate interpretation of the Squantum "tillite" of Massachusetts in light of recent strides in sedimentation, reexamines supposed evidences of ancient glaciation in general, and draws comparisons between the Squantum and Precambrian Gowganda and Utah "tillites." The present study is based chiefly upon comparative analysis of sedimentary structures and stratigraphic characteristics as one approach to re-evaluation of "tillites." The author does not discount categorically all ancient glaciations as such. If uniformitarianism be valid, then probably the earth has experienced more than one frigid episode. But the writer holds that much of the evidence cited is erroneous and must be carefully reviewed.

ACKNOWLEDGMENTS

While stationed at Air Force Cambridge Research Center near Boston from 1956 to 1958, the writer studied the Squantum in some detail in his spare time. A grant from the Wisconsin Alumni Research Foundation enabled him to revisit this area in 1960, to examine the Gowganda and Utah "tillites," and to complete laboratory studies. The writer has benefited from discussions with several Boston geologists, particularly Marland P. Billings and Robert R. Shrock. F. J. Pettijohn, J. C. Crowell, and H. R. Wanless have also contributed useful criticisms. Of special importance were contacts with James E. Wright of the U. S. Geological Survey, who made detailed petrographic studies of the Boston rocks. S. A. Tyler, R. F. Black, and M. P. Billings have read the manuscript. J. Torrie offered helpful suggestions concerning the analysis of variance. R. M. Gates gave valuable petrographic advice and introduced the writer to the Gowganda formation north of Iron Bridge, Ontario. J. G. Koch aided with preparation of illustrations.

GENETIC CRITERIA OF ANCIENT GLACIAL DEPOSITS

Certain areas of South Africa and Australia show substantial field evidence of ancient glaciation (DuToit, 1937; Glaessner and Parkin, 1958). Furthermore, paleotemperature data from fossil brachiopods in Australia (based upon oxygen-isotope studies) indicate sea-water temperatures of only 6.5° C to 8° C in marine shales only 20 feet above probable tillite (Compston, 1960, p. 10, and ms in preparation). Successively warmer temperatures were recorded upward in the shales.[1] But most of the criteria cited for the glacial origin of the Squantum (Sayles, 1914) and for many other supposed tillites can be explained as well by other mechanisms.

The most important alternatives are subaqueous and subaerial gravity sliding and flow of masses of sediment soon after deposition. Churning may destroy stratification and produce a heterogeneous, unsorted till-like pebbly mudstone (Crowell, 1957). Turbidity currents, hydrodynamically distinct from but closely related to mass flow, can deposit repetitious

[1] The writer is indebted to Heinz Lowenstam for bringing this significant information to his attention.

graded sand or silt layers which resemble closely seasonal Pleistocene glacial varves.[2] They may deposit coarser graded clastics as well.

Fairbridge (1947), Crowell and Winterer (1953), Crowell (1957), Pettijohn (1957), and Dott (1959) have reviewed problems of genetic distinctions and have questioned the origin of the Squantum. Newell (1957) has reinterpreted Permian "glacial" boulders in northeast Mexico as due to submarine mass movements figs. 1, 2). Blackwelder (1928; 1930) emphasized superficial similarity of tillites to mudflow deposits, and Van Houten (1957) has reinterpreted the Tertiary Ridgway and Gunnison "tillites" of southwest Colorado as volcanic mudflow deposits. Sayles (1914) debated a subaerial-mudflow origin for the Squantum but rejected it in favor of glaciation and ice-floe rafting.

Large exotic fragments suggest ancient glaciation, but composition alone must be judged

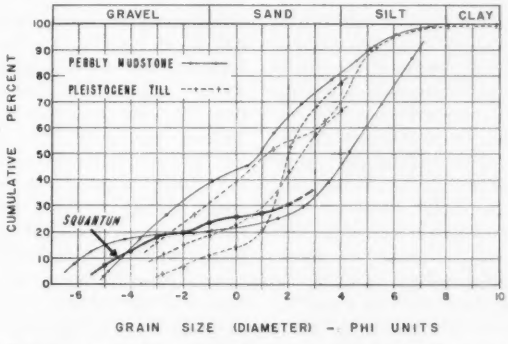

Figure 1. Cumulative size-frequency curves of Pleistocene tills, subaqueous pebbly mudstones, and Squantum "tillite." Because of induration, analysis of Squantum is by volume rather than weight (*See also* Pls. 1, 2).

similar to those that displaced exotic Permian reef-carbonate blocks in Texas (Newell *et al.*, 1953; Rigby, 1958). Sanders (*in* Sanders and Cecioni, 1957) and Zeil (1958) have questioned a glacial interpretation for boulder beds in a Cretaceous geosynclinal sequence of southern Chile possessing some features characteristic of well-known turbidity and mass-flow deposits.

Many "characteristics" of tillites are by no means common in, much less diagnostic of, Pleistocene tills (Flint, 1957). For example, less than 10 per cent of till fragments show striations (Flint, 1957). Faceting and soling are equally elusive. Poor sorting (Pl. 1; Pl. 2, figs. 1, 2), doubtless the most important factor in initial identification of "tillites," can result from quite different agents (Fig. 1; Pl. 1; Pl. 2, cautiously. Pleistocene tills vary greatly in relative proportions of local and foreign debris: some are surprisingly homogeneous (R. F. Black, personal communication). Angular, bent fragments of contemporaneous fine-grained sedimentary rocks in certain pebbly mudstones (Pl. 1, fig. 2) together with contorted zones of interstratified fine sediments suggest gravity movements rather than glaciation. It is virtually inconceivable that fragments of soft, plastic material could survive pressure and grinding in moving ice; however, in a medium of specific gravity and viscosity nearly equal to that of the fragments, their preservation could be expected.

Pettijohn (1957) suggests that isolated pebbles, cobbles, or boulders in fine, evenly laminated sediments constitute strong evidence of ice rafting and the surest indication of frigid climate. Other agents than ice may be important in rafting such debris, however (Crowell, 1957); also, demonstration of ice

[2] Nederlof (1959, p. 643) presents statistical evidence that thickness distribution of seasonal varves may be distinguishable quantitatively from that of thin, graded "turbidites."

rafting does not alone necessitate glaciation—ice rafting can result from seasonal freezing in temperate latitudes. Very abundant, widespread pebbles, cobbles, and boulders can be most readily explained by ice rafting.

The most diagnostic evidence of ancient glaciation is an extensive grooved and polished pavement directly beneath till-like material. Many authors have emphasized the possibility of confusing glacial pavement with structural slickensiding, especially in older, more deformed rocks. Many surfaces are so small as to make interpretation tenuous, notably in eastern Australia (Voisey, 1952). Kenley (1952) and Dunbar (1940) have refuted the suggestions of glacial modification of surfaces in South Australia and Argentina, respectively. Gregory, (1914) reinterpreted a supposed glaciated pavement with gigantic grooves in Peru as a thrust-fault surface. Well-preserved glacial pavements commonly show details such as ice plucking and crisscrossing striations which help to distinguish them from ordinary structural surfaces (Flint, 1957). No pavements are associated with many supposed tillites, including the Squantum. Structure and outcrop character may not be conducive to exposure of pavements in some cases; demonstration of ancient glaciation is greatly impaired in their absence.

THE SQUANTUM "TILLITE"

Introductory Remarks

The writer visited all "tillite" localities and also many exposures of related rocks in and around Boston and studied approximately 100 thin sections. Unfortunately many of the structural relationships and former outcrops are obscured owing to metropolitan growth. Figure 2, a generalized geologic and index map of the Boston area, locates till-like outcrops. Structural relationships have been simplified for illustrative purposes, particularly in the Dorchester-Mattapan area. Several representative partial stratigraphic sections are summarized in Figure 3. For comparison, other deposits appear in Figure 4.

Stratigraphy of the Boston Bay Group

General statement. The Boston Bay group comprises the Roxbury conglomerate, Brighton melaphyre (altered volcanic rocks), Squantum "tillite," and the Cambridge "slate" (argillite). The Mattapan-Lynn volcanic complex underlies the Boston Bay group in most areas, but relationships between the two are not entirely clear. All the units are well indurated, deformed, and locally sheared. The comprehensive reports of Emerson (1917), LaForge (1932), and Billings (1929), the excellent detailed mapping of Billings *et al.* (1939; also unpublished maps), and personal observations have provided the principal sources for the following summary and for the geologic map.

Roxbury conglomerate. The Roxbury conglomerate totals several thousand feet in maximum thickness and consists typically of rounded to subangular pebbles, cobbles, and boulders of quartzitic, volcanic, granitic, and rare syenitic debris with interbeds of sandstone, mudstone (argillite), volcanic tuff, and the lavas and sills of the Brighton melaphyre (Fig. 3; Pl. 2, fig. 3). Volcanic clasts are conspicuously rhyolitic and red, but some are darker andesite and basalt. The different lithologies are not systematically distributed; therefore the Roxbury cannot be subdivided satisfactorily (Billings, 1929). Furthermore, its thickness varies greatly (Billings *et al.*, 1939). In many places the conglomerate is massive and virtually unstratified; however in the Newton area and in Calvary Cemetery, Dorchester (Figs. 2, 3), it is stratified and contains intercalations of appreciable graywacke sandstone and argillite. Shrock (1948) figured many sedimentary features in these interbeds. The sandstones are feldspathic wackes (Williams *et al.*, 1954) containing approximately 40 per cent quartz, 25 per cent siliceous metasedimentary rock fragments (chiefly argillite and quartzite), 15 per cent feldspar (including some microcline), 1–2 per cent volcanic fragments, and about 19 per cent sericitic and quartz-silty matrix.

At Nantasket the conglomerate is rich in volcanic material and contains massive units as much as 100 feet thick with granite boulders up to 8 feet in diameter mixed with volcanic pebbles and cobbles averaging about 3 inches in diameter (Fig. 3). Here volcanic grains constitute at least 25 per cent of sand-size detritus, producing a volcanic wacke matrix (Williams *et al.*, 1954). Many of the coarse fragments at this and some other localities are neither well rounded nor even moderately well sorted. The Roxbury deposits show characteristics of rapid accumulation with only modest transport and abrasion. In composition the Roxbury is an immature, polymictic conglomerate.

Cambridge "slate". The Cambridge "slate," better termed argillite, consists of 2000–3000 feet of typically dark, laminated tuffaceous

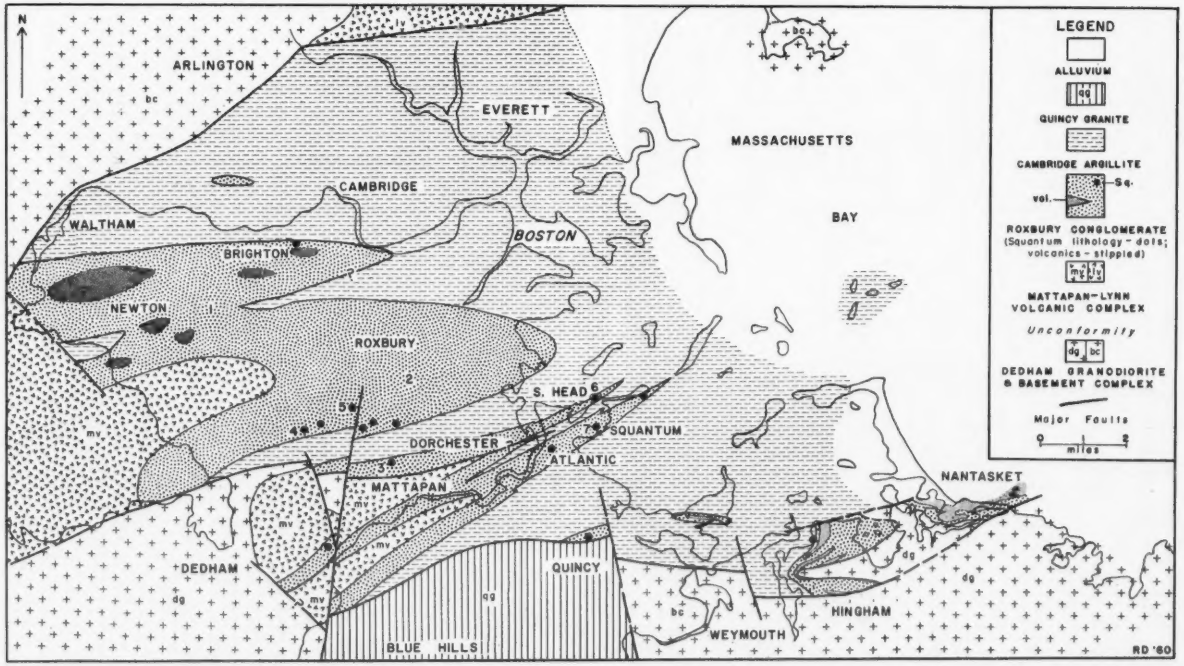

Figure 2. Generalized geologic and index map of the Boston area. Numbers designate localities where pebble counts were made.

siltstone and mudstone with sandstone or quartzite at the top. The formation is well indurated and locally displays a fracture cleavage, but it is not a true slate (Billings, 1929). Excellent examples of intricate sedimentary features such as small-scale current stratification, scour marks, load casts, and some graded bedding characterize the Cambridge; numerous contorted zones range from a few inches to several feet in thickness (Fig. 3; Pls. 3, 4; Shrock, 1948). Exposures are rare, and most are extremely poor. The mineralogy includes quartz, sericite, feldspar, melaphyre fragments, chlorite, and epidote, which suggest igneous sources (LaForge, 1932). Apparently some of the detritus was pyroclastic, and several tuff beds have been reported (LaForge, 1932). In the hope of identifying clay minerals which might be indicative of the environment of deposition, the author had X-ray analyses made of two samples. S. W. Bailey reported that only $2M_1$ muscovite was present. Probably it resulted from recrystallization due to low-grade metamorphism.

Squantum "tillite." The Squantum is a distinctive heterogeneous and very poorly sorted rock consisting of rare boulders up to 4 feet in diameter with cobbles, pebbles, sand, and a large percentage of fine matrix of silt- and clay-size particles (Fig. 3; Pl. 1, fig. 1; Pl. 2, fig. 1). Matrix constitutes as much as 50–63 per cent by volume. Figure 2 indicates localities where Squantum lithology has been reported. The Atlantic locality has been very well exposed by Pleistocene glacial abrasion; detailed relationships of Roxbury, Squantum, and Cambridge lithologies are nowhere better revealed. Squantum sand- and gravel-size detritus has essentially the same composition as that of the Roxbury, *i.e.*, it has volcanic, granitic, metasedimentary, feldspathic, and quartzose fragments in approximately the same ratios; it is an excellent example of a lithic wacke. But large angular fragments of laminated argillite identical with that of the Cambridge constitute a very important peculiarity of the "tillite." Some of these penecontemporaneous fragments are bent and deformed, indicating clearly that they were soft and plastic when redeposited (Pl. 3, fig. 1; *Compare with* Pl. 1, fig. 2). Emerson (1917) interpreted the Squantum as an upper member of the Roxbury, but Billings (1929) distinguished it as a separate formation. For reasons discussed subsequently, the present writer does not regard it as a separate stratigraphic unit. Thickness ranges from zero to a few hundred feet. Commonly some well-stratified Cambridge-like laminated argillite, and in some places Roxbury-like conglomerate, are intercalated within or between the "tillite" bodies; therefore quoted thickness figures are misleading (Fig. 3).

Associated volcanic rocks. Volcanic rocks constitute a very important part of the Boston sequence. The Mattapan-Lynn complex, which in recent literature has been regarded as wholly

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PLATE 1. PEBBLY MUDSTONES IN SQUANTUM "TILLITE," SQUANTUM, MASSACHUSETTS, AND AT CAPE BLANCO, OREGON

Figure 1. Typical outcrop of heterogeneous, unsorted Squantum "tillite," northeast end of Squantum Head, Squantum, Massachusetts. Note bent mudstone mass just above and left of center and rounded character of other clasts.

Figure 2. Unsorted, unstratified marine pebbly to bouldery mudstone within a stratified, graded sandstone-mudstone sequence, Cape Blanco, Oregon (Jurassic or Cretaceous). The pebbly mudstone is 60 feet thick (maximum) and is broadly lenticular. Note penecontemporaneous laminated mudstone-sandstone blocks.

PLATE 2. PHOTOMICROGRAPHS OF PEBBLY MUDSTONES AT ATLANTIC LOCALITY, MASSACHUSETTS, AND CAPE BLANCO, OREGON, AND OUTCROP OF ROXBURY CONGLOMERATE, MASSACHUSETTS

Figure 1. Photomicrograph of the finer phase of the Squantum "tillite," Atlantic locality. Extremely poor sorting, abundance of angular quartz detritus, and fine, submicroscopic matrix are characteristic. Crossed nicols, 5 ×

Figure 2. Photomicrograph of the finer phase of the Cape Blanco, Oregon, pebbly mudstone and sandstone. Very poor sorting and heterogeneity of composition are similar to the Squantum, although average grain size is slightly greater. Crossed nicols, 5 ×

Figure 3. Typical Roxbury conglomerate outcrop, Atlantic locality, stratigraphically below area shown in Figure 3 of Plate 3. Subrounded volcanic, sedimentary, and plutonic rock fragments are represented with variable proportions of quartzose sand matrix.

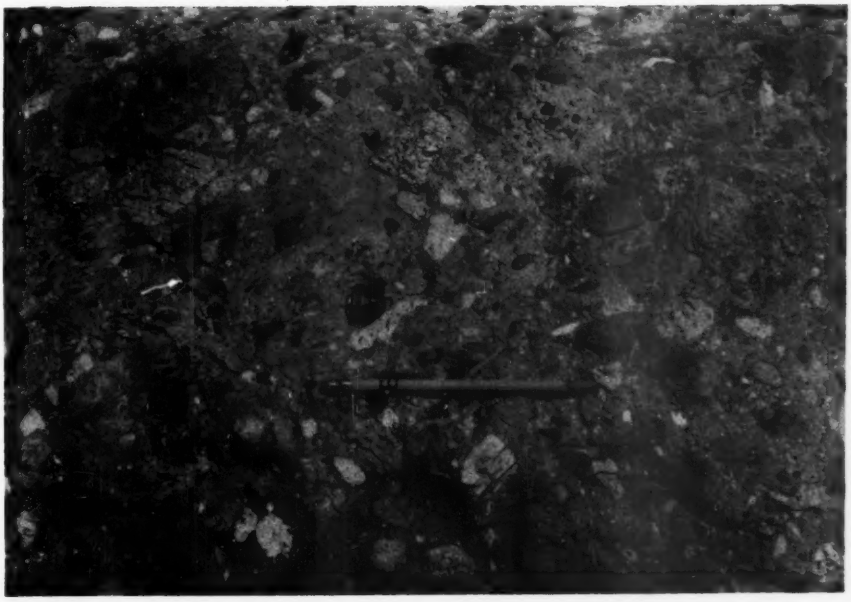

Figure 1

Figure 2

PEBBLY MUDSTONES IN SQUANTUM "TILLITE," SQUANTUM, MASSACHUSETTS, AND AT CAPE BLANCO, OREGON

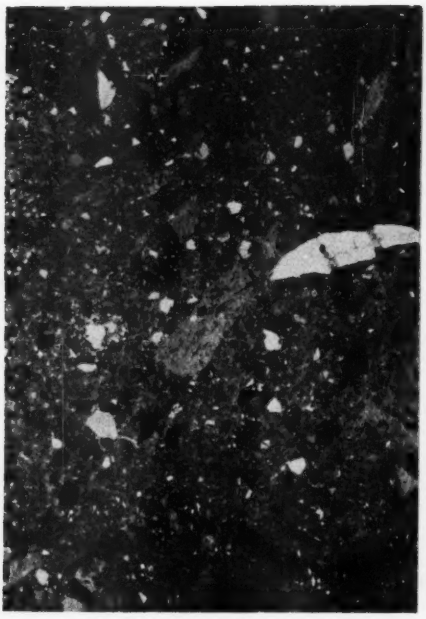

Figure 1

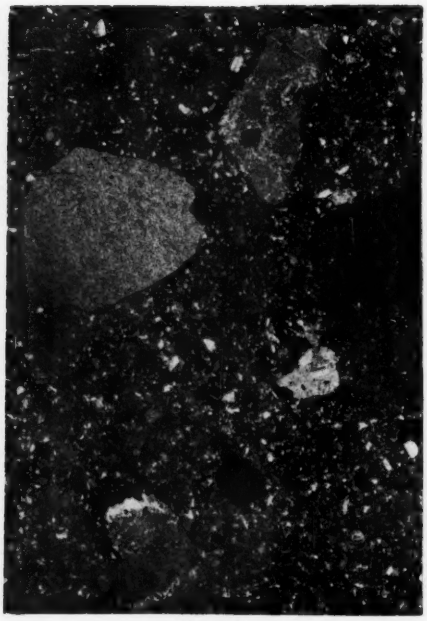

Figure 2

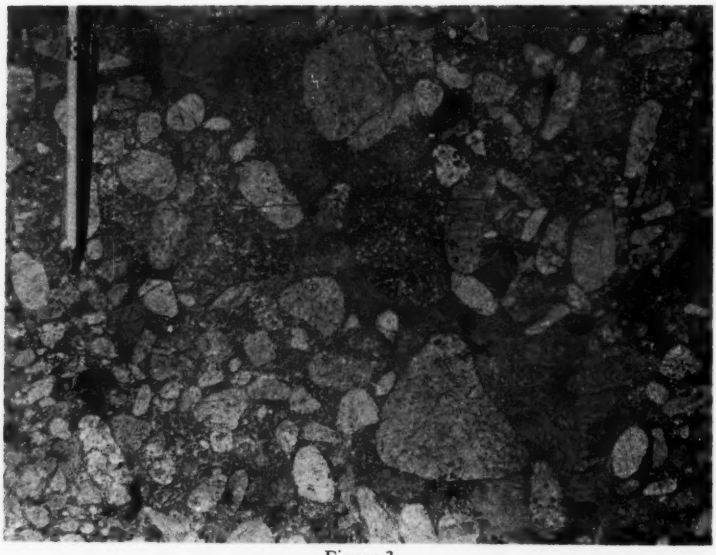

Figure 3

PHOTOMICROGRAPHS OF PEBBLY MUDSTONES AT ATLANTIC LOCALITY, MASSACHUSETTS, AND CAPE BLANCO, OREGON, AND OUTCROP OF ROXBURY CONGLOMERATE, MASSACHUSETTS

Figure 1

Figure 2

Figure 3

CONTORTED STRATIFICATION IN VARVELIKE ARGILLITE,
ATLANTIC LOCALITY, MASSACHUSETTS

Figure 3

Figure 2

Figure 1

GRADED BEDDING IN TWO LOCALITIES NEAR SQUANTUM, MASSACHUSETTS, AND IN GOWGANDA FORMATION, ONTARIO

pre-Roxbury, contains red to tan rhyolitic, andesitic, basaltic and local keratophyre flows, coarse granitic and syenitic dikes, volcanic breccias, scoria, and volcanic mudflows (LaForge, 1932). Thickness appears to exceed 1000 feet. Emerson (1917) described supposed volcanic-vent or explosion breccias composed of volcanic and granitic blocks in West Roxbury and Hyde Park (between Dedham and Mattapan, Fig. 2), and this writer observed local coarse, red cataclastic granite dikelike masses in Needham and Dorchester. Pebbles of identical lithology are present in the Roxbury. Presumably related sodic volcanic rocks and dikes are associated with the Quincy granite in the Blue Hills area (LaForge, 1932, p. 35). The Brighton melaphyre and related volcanic rocks are interstratified rather widely within the Roxbury and consist of rhyolitic to basaltic flows, dikes, breccias, and tuffs.

Stratigraphic relationships. Published interpretations of the Boston Bay succession have presented the Roxbury as the lowest in a simple, vertical sequence resting unconformably upon the presumably older Mattapan-Lynn complex or upon the Dedham granodiorite, depending upon locality (LaForge, 1932; Billings, 1929; Billings *et al.*, 1939). The Squantum and Cambridge were thought to overlie the Roxbury successively and to be distinct from it in spite of published descriptions of interstratified "Cambridge-like strata" beneath and within the Squantum and intercalated at several places within the Roxbury (Fig. 3) as well as in the supposedly older Mattapan volcanic rocks (LaForge, 1932). Early workers (Emerson, 1917; LaForge, 1932) did not regard any of the Cambridge-like strata as belonging to the Cambridge but considered each a separate rock unit or a member of the Roxbury and coined local names for them (Dorchester slate, *etc.*). Some are lighter-colored than the typical Cambridge, notably at Squantum Head where they are pink, but otherwise are lithologically identical. Taken in historical context, such conclusions are not surprising, for concepts of sedimentary facies had not yet been clearly formulated and widely accepted. Simple superposition of these units seems consistent with mapping of the areal geology (Fig. 2), but the writer believes that at least some intertonguing of various lithofacies exists in the Boston sequence even though such a relationship cannot be proved satisfactorily owing to poor exposures and structural complications. Billings *et al.* (1939) have shown clearly that intricate facies variations do exist at Hingham and Nantasket between clastic and volcanic lithologies within the Roxbury (Figs. 2, 3). The writer noted also in the field that interstratification of sandstone, argillite, and till-like lithologies within the Roxbury conglomerates becomes more common toward the northeast. This could result solely from fortuitous outcrops or structural position; however the contrast between sequences around Rox-

PLATE 3. CONTORTED STRATIFICATION IN VARVELIKE ARGILLITE, ATLANTIC LOCALITY, MASSACHUSETTS

Figure 1. Large broken and bent fragments of Cambridge argillite lithology in "tillite," Atlantic locality, 15 feet above Figure 2 of this plate. The fragments must have been soft when torn up and incorporated in the "tillite."

Figure 2. Laminated argillite with erratic pebbles and pebbly zones overlain by poorly sorted conglomerate or "tillite," Atlantic locality. Gravity loading along the contact apparently was arrested just before renewed large-scale mass flow could set in.

Figure 3. Large-scale contorted zone in argillite intercalated within the upper Roxbury conglomerate, Atlantic locality, approximately 100 feet below Figure 2 of this plate. "Tillite" occurs above the argillite to the right of view. Note the irregular, gravity-loaded contact with underlying conglomerate.

PLATE 4. GRADED BEDDING IN TWO LOCALITIES NEAR SQUANTUM, MASSACHUSETTS, AND IN GOWGANDA FORMATION, ONTARIO

Figure 1. Graded pebbles and sand, top of the Squantum "tillite" along the south shore of Squantum, Massachusetts. This zone and others southeast of Squantum Head are interpreted as material settled from slurries produced just above gravity mass flows.

Figure 2. Crudely graded silt and mustone, argillite lithology, Squantum Head. Graded bedding in the argillites is more variable in scale than is typical of true varves and of the Gowganda formation. Crossed nicols, 2.5 ×

Figure 3. Supposed varves, Gowganda formation (Precambrian) near Endikai Lake, Ontario. Upper laminae are truly varvelike, but lower graded silt seems abnormally thick. Crossed nicols, 2.5 ×

bury and Arnold Arboretum and those at Atlantic and Squantum are striking. Finally, the writer's experience with similar thick, coarse conglomeratic sequences in Wyoming (Eocene), Nevada (Pennsylvanian), the Ventura basin (Miocene and Pliocene) and southwest Oregon (Cretaceous and Eocene) indicate that such coarse deposits tend to grade abruptly (within a relatively few miles at most) into finer sediments normal to regional depositional strike, *i.e.*, away from actively rising tectonic or volcanic lands.

The writer believes even more emphatically than did Sayles (1914) that complete lithologic gradations exist between typical Roxbury and typical Squantum rocks. In the writer's view, then, the Squantum simply constitutes lenses of more poorly sorted Roxbury material together with incorporated angular slabs of Cambridge lithology. Exact stratigraphic position of individual Squantum outcrops is indeterminate, but it is very unlikely that they are all laterally continuous or necessarily even at the same stratigraphic position. In Mattapan, "tillite" appears to be near the base of the Roxbury, whereas elsewhere it is near the top. In the town of Squantum, three and possibly more separate zones are present. Two occur at Squantum Head (Fig. 3), and another thick one is found in the southeast part of the town of Squantum with pink argillite and normal Roxbury lithologies intervening (Fig. 2). Past writers have attributed such alternations solely to fault repetition both at Squantum and around Mattapan and Dorchester. Because undisputed interstratification is present locally, the writer suggests that interstratification may occur on a larger scale as well, even though some of the relationships are complicated by faulting. Billings (Personal communication), however, does not agree with this interpretation. It is unfortunate that no stratigraphic datum exists in the sequence which would provide undisputed lateral correlation to test the facies hypothesis, for it has considerable import in the interpretation of conditions of deposition and paleogeography.

Emerson and LaForge cite evidence suggesting genetic relationship and gradation between the Boston Bay group and the Mattapan volcanic rocks. Emerson concluded that volcanism began before ("Mattapan"), but continued essentially without interruption during Roxbury deposition ("Brighton"). Absence of Mattapan volcanic rocks between Roxbury-like conglomerate and the Paleozoic or Precambrian Dedham granodiorite in Hingham and Nantasket (Fig. 2), together with abundance of Mattapan debris in the conglomerate, could constitute evidence of a major unconformity between the Roxbury and Mattapan. However, the discontinuous Mattapan distribution may only reflect geographic variation in original volume of lavas extruded in an essentially continuous stratigraphic succession which was characterized by progressive transgressive onlap at its base. Such variability during Roxbury deposition is borne out by studies of Billings *et al.* (1939), which showed also a profound unconformable onlap of the Boston Bay group over a surface of considerable relief developed on the Dedham granodiorite in Hingham (a "buttress unconformity" of California geologists).

Age of the Boston Bay Group

The age of the Boston Bay group remains in question, yet is quite pertinent to the glacial hypothesis. Two poorly preserved casts of presumed fossil trees which lacked any bark impressions or internal structure were found in the Roxbury (Mansfield, 1906) and subsequently lost. No other fossils have ever been reported. Elso S. Barghoorn recently rediscovered one of the tree trunk casts at Harvard. He identified it (Personal communiation) as possibly *Callixylon* or *Cordaites*, but definitely not *Lepidodendron* or *Sigillaria*. Barghoorn states that it could even be a post-Paleozoic plant, but because all other evidence indicates that the sequence is almost certainly Paleozoic, this fossil is thought to range between late Devonian and late Permian. Attempts to extract microfossils have been unsuccessful. Superficial lithologic resemblance to Pennsylvanian coal-bearing strata of Rhode Island led to early assignment of a late Paleozoic age for the Boston rocks. It is apparent from older literature that circular reasoning played a role in establishing this age assignment: first the two fossil trees and supposed lithologic similarity with Upper Carboniferous Narragansett strata of Rhode Island "established" a late Paleozoic age; then presence of "glacial tillite" and the recognition of widespread Permian glaciation in every continent except North America pointed to a Permian age for the Squantum (Sayles and LaForge, 1910; Sayles, 1914); subsequently the "established" Permian age was invoked to fortify the glacial interpretation.

LaForge (1932) and Shrock (Personal communication) have pointed out that supposed

similarity of the Boston and Rhode Island strata is unconvincing. The Rhode Island deposits are prominently bright red in part, contain abundant plant material (including coal), and unconformably overlap and contain detritus of a blue-gray sodic, mica-poor, hornblende granite similar to the Quincy granite of eastern Massachusetts which is not reported in Roxbury or Squantum fragments (Mansfield, 1906; LaForge, 1932; Quinn, 1953). Granitic debris in the Boston Bay strata is of the supposed Precambrian or early Paleozoic Dedham type, *i.e.*, white to pink granodiorite containing plagioclase, some microcline and biotite, but little or no amphibole or pyroxene (LaForge, 1932; Billings *et al.*, 1939). There are only very minor associated volcanic rocks in the Rhode Island sequence (Quinn, 1953; Towe, 1959), and it possesses different stratification characteristics. Also significant are paleocurrent data which indicate that the Rhode Island sediments were transported generally from north to south and apparently were derived from erosion of a metamorphic, sedimentary, and plutonic terrain in the Boston area itself (Towe, 1959). For these reasons the writer agrees with LaForge's and Shrock's belief that the two sequences are not equivalent. Reported absence of Quincy detritus from the Boston Bay rocks suggests that the Boston Bay sequence is older, and is also older than the Rhode Island deposits. However, Emerson (1917) and LaForge (1932) suggested that the Mattapan volcanic rocks together with dikes that crosscut them were chemically akin to the Quincy and essentially contemporaneous. Therefore LaForge (1932) considered that the Quincy pre-dated the Roxbury and that it was not unroofed by erosion until after deposition of the Cambridge argillite.

Webber *et al.* (1956) dated, by the lead-alpha method, zircons from granite considered correlative with the Dedham at 355–365 million years, or late Devonian (Acadian) on the new Kulp scale (Kulp, 1960). A K-Ar whole rock date for the type Quincy granite has yielded a *minimum* age of 280±15 million years, and a younger *minimum* age of 255 million years has been determined for the overlying early Pennsylvanian Narragansett group on the basis of K-Ar and Rb-Sr measurements of metamorphic micas (Hurley *et al.*, 1960). This together with the other evidence suggests that the Boston Bay group and Quincy granite are neither Pennsylvanian nor Permian, but probably Mississippian or Devonian.

Alternate Hypothesis for Origin of the Squantum Deposits

Reasoning about the Squantum must be to a large extent by analogy with more easily interpreted deposits. The Boston Bay group and the Paleozoic sedimentary and volcanic rocks and granites throughout New England and the Maritime Provinces of Canada possess many attributes in common with younger sequences in similar tectonic belts such as those of the Pacific Coast (Figs. 3, 4). All lie within old eugeosynclines which had long histories of more or less continuous volcanism and orogeny. Presence of abundant volcanic rocks of diverse sorts in complex association with clastic sediments at Boston is typical of such provinces.

The volcanic and coarsely clastic character of the Boston Bay group points to deposition associated with an orogenic pulse that may have achieved full development with emplacement of the Quincy granite. Definite nonmarine, coal-bearing Pennsylvanian strata of Rhode Island represent separate, "postorogenic" deposits resulting from later erosion of the tectonic lands raised within the geosyncline, as do equivalent deposits of the Maritime Provinces (Kay, 1951). The Rhode Island sediments were themselves folded, metamorphosed and intruded locally by granites after the Pennsylvanian (Quinn, 1953; Hurley *et al.*, 1960).

Environments in which the Boston sedimentary and volcanic rocks were deposited are almost as suspect as their ages. The fine lamination, current structures, and graded bedding in the Cambridge and limited parts of the Roxbury lithologies indicate deposition both by currents and by gravity settling in an aqueous medium of unknown depth. But there is no present basis for proving whether this was lacustrine or marine. A transgressive marine setting was envisioned by all workers prior to 1875 when Dodge first suggested a glacial origin for the Squantum, but once the glacial hypothesis became well established, a nonmarine setting was deemed necessary for the whole sequence. Such is not in accord, however, with the paucity of sedimentary characters typical of such environments, *i.e.*, abundant scour channels, extensive but erratically oriented, large-scale cross-stratification, great abundance of terrigenous fossils to the exclusion of marine ones, coal, and the like. By contrast, these features are all plentiful in the Rhode Island Pennsylvanian strata. A periglacial lake has been envisioned as the most

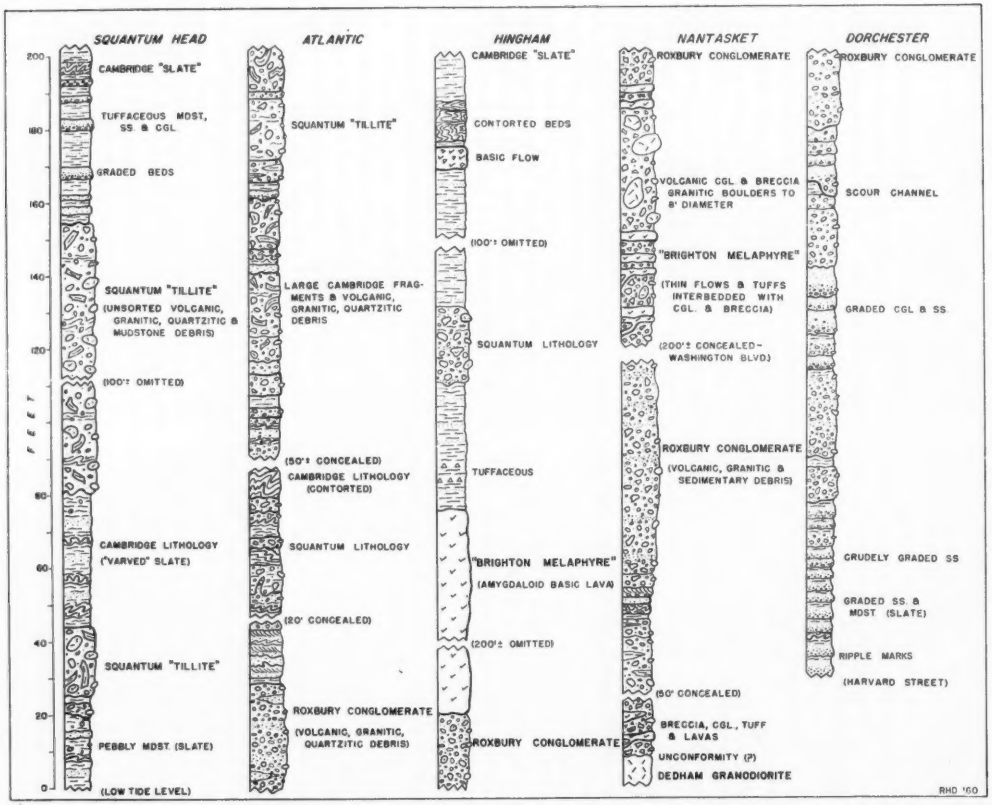

Figure 3. Representative sections of the Boston Bay group showing lithology and stratigraphic relationships of the Squantum "tillite." *See* Figure 2 for location of sections. (All sections described by the writer)

Figure 3. Representative sections of the Boston Bay group and of the Squantum "tillite." *See* Figure 2 for location of sections. (All sections described by the writer)

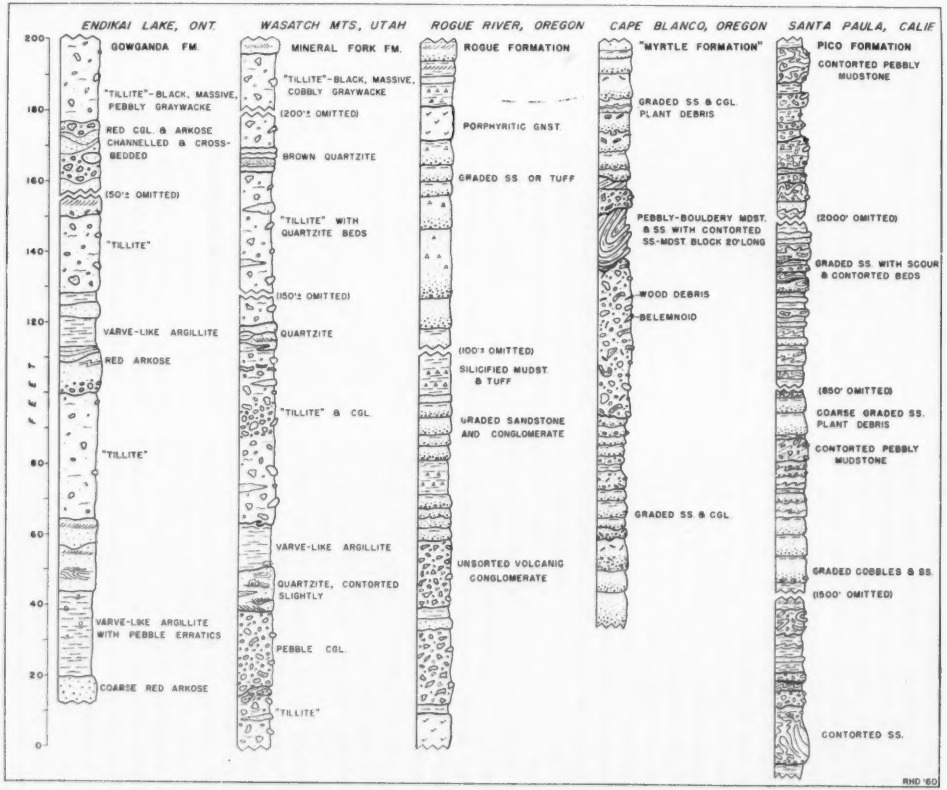

Figure 4. Deposits possessing features in common with the Boston Bay group. Two columns at the left are of uncertain origin and are Precambrian; remaining three are marine, nonglacial and Mesozoic or Cenozoic. (All sections described by the writer; thicknesses approximate)

probable site of deposition for the "varved" Cambridge argillite (Sayles, 1916; Lahee, 1914a), but this, too, was based upon a glacial premise. The Roxbury was considered to have formed from outwash gravel and sand, and periodic advances of valley or piedmont glaciers across these deposits were supposed to have laid down till which is now intercalated within Roxbury and Cambridge strata. Advance of the ice was thought to have deformed the fine lake sediments locally, although Lahee (1914a; 1914b) suggested grounding of icebergs as the cause of contorted argillite zones. Iceberg rafting was invoked here, as elsewhere, to explain rare anomalous coarse erratic fragments embedded in fine, laminated mudstone or argillite (Fig. 3). Thus an apparently complete and consistent restoration emerged, and it is readily understandable that the hypothesis of glaciation has gone unchallenged so long. The writer tentatively favors a marine environment for the better stratified sedimentary rocks and for the Squantum by analogy with similar eugeosynclinal strata elsewhere (Figs. 3, 4). He interprets most of the Roxbury conglomerate and associated volcanic rocks as nonmarine or very near-shore deposits laid down adjacent to a volcanically and tectonically active land or islands. Presence of pyroclastic material and the thickness and coarseness of the Roxbury conglomerate clearly indicate presence nearby of significant land areas, although size and exact position cannot be determined accurately. Clearly Mattapan, Dedham, Brighton, and other rocks all were being actively eroded nearby during a period of considerable tectonism characterized by progressively diminishing volcanism.

All known detritus in the Squantum seemingly can be readily accounted for in the immediate Boston Bay area (Sayles, 1914). But as a test of the subaqueous gravity movement hypothesis of Squantum deposition, the writer made random counts of pebbles larger than 1 inch in diameter at 10 widely scattered localities—five each of Roxbury and Squantum lithologies. Angular fragments of Cambridge argillite in the Squantum were omitted, as their peculiar derivation is obvious and irrelevant to this particular test of resedimentation. Fragments larger than sand-size were chosen because coarser debris constitutes the most sensitive indicator of immediate source (fine material is more likely a result of multiple cycles of redeposition). Comparison with point-count data for associated sandstones verified this conclusion. Data from three localities were omitted from later statistical analysis (Table 1) because they represented rather atypical local volcanic conglomerate and "tillite" readily accounted for by their stratigraphic proximity to volcanic rocks in Dorchester and Milton. For the remaining data, from apparently more homogeneous localities, an analysis of variance was employed to determine if the variance of percentages of sedimentary, volcanic, and granitic fragments between the Roxbury and

TABLE 1. PEBBLE-COUNT DATA USED IN ANALYSIS OF VARIANCE OF ROXBURY AND SQUANTUM LITHOLOGIES

See Figure 2 for locality distribution. $F = 2.31$, not significant, ($F_{.05(2,10)} = 4.10$)

Localities	Sedimentary*	Volcanic	Granitic
Roxbury conglomerate			
1. Chestnut Hill	59	33	8
2. Franklin Park	55	38	7
3. Harvard Street	55	42	3
Subtotals	169	113	18
Squantum "tillite"			
4. Weld Street	55	35	10
5. Arnold Arboretum	50	39	11
6. Squantum Head	43	48	9
7. Southeast Squantum	51	43	6
Subtotals	199	165	36

* Includes quartzite and siliceous argillite fragments

Squantum was greater than variance of the percentages between localities within each. The value obtained by the standard "F" test (Steel and Torrie, 1960) indicated no significant difference for these data (Table 1). The writer takes the close similarity of Roxbury and Squantum clast compositions, except for the contemporaneous Cambridge argillite fragments in the Squantum, to indicate that the "tillite" could well have been derived simply by redeposition of Roxbury gravels and mixing with finer muds. This alone does not disprove glaciation, but if the compositions differed significantly, there would be stronger necessity for postulating ice transport to account for materials that could only have come from far away, particularly those of large size.

The writer believes that Squantum fragments are not greatly different in shape from those of the Roxbury either (Pl. 1, fig. 1; Pl. 2, fig. 3), and that later shearing, particularly at Squantum Head, has accentuated their angular appearance in outcrop. Both faceting and striation of clasts are exceedingly rare in the

Squantum (Sayles, 1914; Lahee, 1914a). The writer found none.

Numerous intensely contorted zones from a few inches to 10 feet thick lying between completely undisturbed strata (Figs. 2, 3; Pl. 3, figs. 2, 3) are best explained by penecontemporaneous plastic mass flow on a sloping surface. Shrock (1948) suggested this origin for some of the small contorted zones in the Boston Bay group, but not for the Squantum itself. Figure 2 of Plate 3 illustrates moderate deformation due to loading, and Figure 3 shows much larger-scale contortion apparently due to considerable lateral flow. The unsorted Squantum lithology (Pl. 1, fig. 1) represents still greater churning and more extensive flow. Intimate relationship of Squantum and Roxbury lithologies and the possible intertonguing of the two together with graded bedding in various lithologies (Fig. 3; Pl. 4, figs. 1, 2) convinced the writer that mass flow and local turbidity-current deposition of clastic sediments in a tectonically and volcanically active region presented more plausible mechanisms without necessity of invoking one of the most unusual phenomena known, glaciation. True seasonal varves are characteristic of fine sand, silt, and clay, whereas turbidity-current deposits may include graded cobble, pebble, and coarse-sand units several feet thick as well as much finer and thinner graded units (Fig. 4). Such extreme associations are found at Boston as illustrated in Figure 3, and Figures 1 and 2 of Plate 4. Intercalated stratified zones in the upper part of some Squantum units (Pl. 4, fig. 1), cited by Sayles (1914) as a criterion of tillites, are best explained as representing the first resumption of more normal subaqueous deposition after cessation of gravity movement. Most of these zones apparently represent simple graded gravity settling of gravel and sand slurries formed in immediately overlying water by movement of the flows themselves, and are, strictly, distinct from true turbidity-current deposits. Jones (1937) and Lippert (1937) noted this type of deposition in submarine gravity deposits. Rapid dumping of gravels along a shore-line zone adjacent to volcanic lands, which were doubtless seismically active as well, could have readily produced conditions favorable for sliding, mass flow, and mixing of gravels with finer, offshore muds, and silts.

The writer envisions normal currents moving fine, in part pyroclastic, material in unknown depths of water with occasional incursions of subaqueous slide, mass flow and flow-induced turbidity currents which introduced coarser gravel, sand, and silt. Turbidity currents cannot be proven to have been the major factor in deposition of the Cambridge sediments, although probably they were more important than is evidenced today. Much graded bedding may have been destroyed by reworking. The writer is preparing a manuscript on the important distinctions of types of subaqueous gravity deposits.

Subaerial mudflow has been suggested as a possible mechanism for deposition of the Squantum (Sayles, 1914) and for the "volcanic mudflows" in the Mattapan volcanic rocks (LaForge, 1932). The very thick and extensive, finely laminated Cambridge argillite and parallel stratification in some of the Roxbury indicate subaqueous deposition, however. Mudflows and landslides do rarely flow into water from shore areas, as along the Netherlands coast (Terzaghi, 1950). But the Squantum lithology at several localities occurs within well-stratified sequences which seemingly were deposited some distance from shore.

CONCLUSIONS AND PALEOGEOLOGICAL COMPARISONS WITH GOWGANDA AND UTAH TILLITES

The Boston Bay group is reinterpreted here as having formed from complex lithofacies of marine or lacustrine as well as subaerial gravels, sands, laminated mud and silt, lavas, and pyroclastics deposited adjacent to volcanic highlands. Interstratified Squantum "tillite" lenses bear so many features in common with well-established subaqueous gravity-mass–flow deposits that, by analogy, a similar origin seems more plausible than glaciation. Moreover, probability of a pre-Pennsylvanian age weakens the argument for glacial origin.

Poor sorting, lack of stratification, and most other supposed criteria of tillites are in reality undiagnostic. The most convincing evidence, a widespread gouged and polished pavement beneath till-like deposits in clearly nonmarine strata, is lacking at Boston and in many other places. Nearly horizontal striated and polished surfaces do appear locally beneath the Gowganda tillite in Quebec (Cooke, 1923) and in eastern Ontario (Paul E. Schenk, Personal communication[3]). The Squantum possesses neither significant numbers of erratics in laminated mudstones nor any rock fragments

[3] Graduate student, University of Wisconsin. Schenk discovered two exposures of a nearly horizontal striated surface beneath the Gowganda in Vogt Township southeast of Lake Temagami, Ontario, in 1960.

that could not have been derived locally. Unlike the Gowganda and some of the better-established tillites of the southern hemisphere, all the debris in the Squantum could have originated merely by redeposition of local gravels with addition of torn-up and bent fragments of contemporaneous soft muds.

Furthermore, volume of normal (*i.e.*, better sorted) conglomerate, which was the postulated source of the Squantum, greatly exceeds "tillite" at Boston, as should be expected if the "tillite" originated by gravity flow and sliding of unconsolidated gravels. But the Gowganda and Utah unsorted "tillites" greatly exceed associated sorted conglomerate in volume. If normal gravel was the source of these, too, by gravity movements, then practically all of it either was redeposited to form unsorted "tillite"—which seems unlikely—or else it has been removed by subsequent erosion. A glacial origin, however, is not incompatible with the preponderance of "tillite" over conglomerate.

The Squantum, Gowganda, and Utah "tillites" contain a large proportion of rounded pebbles and boulders; the associated lithologies also reveal considerable current action (Figs. 3, 4). The Utah "tillite" also contains unusual numbers of well-rounded quartz-sand-size grains in certain zones, whereas most quartz-sand-size grains in the Gowganda and Squantum are angular. Both the Utah and Gowganda "tillites," on the other hand, are characterized by a conspicuously greater percentage of sand and clay matrix than the Squantum. Recognition of graded bedding in siltstone, sandstone, and conglomerates of the Boston Bay group suggests that some of the laminations in the finer sediments resulted from turbidity currents rather than a strictly seasonal varve phenomenon. The Gowganda shows only fine, thin, more truly varvelike graded bedding (Pl. 4, fig. 3). Gowganda erratics, some very large, are so abundant and widespread as to strongly suggest ice rafting. The writer did not observe definite graded bedding and erratics at either the Mineral Fork or Rock Canyon localities, Utah, although they may be present. Blackwelder (1932) reports varvelike argillites in these deposits.

Independent cold-climate indicators such as diagnostic plants or animals and isotopic paleotemperature data would support ancient glaciation, but they are lacking at Boston and in most other "tillite" localities. Even the famous *Glossopteris* flora associated with southern hemisphere, late Paleozoic tillites has never been proven botanically to be a cold-climate flora (DuToit, 1937, p. 83).

In the final analysis, over-all paleogeological relationships must constitute essential—probably the most essential—factors in evaluation of evidence for ancient glaciation. The local extent, position within an old orogenic belt, abundance of volcanic material, probable close relationship to major orogenic episodes, and striking similarity of over-all character with deposits of other eugeosynclinal belts compel the writer to reject the glacial hypothesis for the Squantum. This conviction is reinforced by the enigma of a single, small, isolated "glaciated" area in North America where general paleogeographic evidence suggests contradictory climatic conditions. Thickness and mature quartz arenite-limestone lithologies suggest that the Bruce-Cobalt sequence containing the Gowganda (Collins, 1925; McDowell, 1957) and the Cottonwood series containing the Mineral Forks tillite, Utah (Crittenden *et al.*, 1952), were, by contrast, probably both deposited at cratonic-miogeosynclinal margins. The Ontario sequence thickens and becomes finer southeastward, and the Utah sequence thickens westward. Both lack contemporaneous volcanic rocks. These "tillites" are far more extensive than the Squantum (Collins, 1925; Cohenour, 1959), yet poorly sorted, immature graywackes *as thick and extensive* as these are virtually unknown in strata of analogous younger cratonic margins. Present modest structural attitudes and broad disconformable overlap at the base of the Gowganda indicate a moderately subdued ancient terrain except locally (Collins, 1925; Cooke, 1923). On the other hand debris in the Gowganda ranges up to 15 feet long and is overwhelmingly of much older, pre-Bruce series rocks seemingly transported great distances, at least in western Ontario. Great relief, as in a continuous mountain chain, would seem indicated to account for simple gravity movement of so much old, coarse debris over a region at least 300 by 100 miles (probably much greater). Known gravity deposits tend to be far more limited areally. Furthermore, large-scale contorted zones and contemporaneous laminated argillite fragments are not associated with either the Gowganda or Utah examples, although cross bedding, scours, and small contorted zones are known in current-deposited sandstone and argillite intercalations (Fig. 4). Regardless of origin of these "tillites," associated deposits attest to very complex, fluctuating environments and

processes. In absence of certain features characteristically associated with subaqueous mass flow deposits, it appears plausible that continental glaciers and icebergs that extended into marginal marine areas may, after all, satisfactorily explain both the Gowganda and nearly identical Utah "tillites." Even if this is correct, however, there can be no assurance of their synchroneity.

The proposed tillites of South Africa, India, western and central Australia, and southeastern Brazil all lie in stable, nonorogenic, cratonic regions and are contained in chiefly nonmarine sequences. They are not severely deformed and are geographically extensive. In contrast, those of eastern Australia and Tasmania, western Argentina, Chile, and Bolivia as well as the Eo-Cambrian "tillites" of Scandinavia all lie within orogenic belts and have associated rocks typical of unstable geosynclines. Of all these examples, a glacial origin is best established for the deposits in the cratonic areas. While several of the examples from ancient unstable orogenic belts probably also are glacial, notably the proposed tillites in Australia and Tasmania, gravity cannot be discounted.

The famous North American Levis (Quebec), Johns Valley (Oklahoma), and Haymond (southwest Texas) boulder erratics in dark, marine mudstone sequences, which have in part been attributed to ice rafting, also lie in orogenic belts and are associated with zones of overthrusting. The writer regards a subaqueous gravity origin as far more compatible with their paleogeological settings.

Promiscuous postulation of ancient glacial periods still exercises imaginations of scientists; however most supposed examples of ancient glacial deposits must be critically re-examined. Many others may have been formed by one of the several possible alternate mechanisms. To be valid, postulated glaciations must be compatible with paleogeographic and paleotectonic evidence. Theories of geotectonics, paleoclimatology, and paleobiogeography based wholly or in part upon supposed ancient glaciations, including continental drift and polar wandering, cannot be evaluated honestly until evidence for all such periods has been critically reanalyzed. If the present paper does nothing else, it underscores the great difficulty in interpreting glacial-like deposits.

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Origin of the Gulf of California

Abstract: The probable cumulative Late Cretaceous and Cenozoic right-lateral strike-slip displacement along the San Andreas fault in central California is 350 miles. The San Andreas and the allied faults into which it branches southward trend longitudinally into the Gulf of California, and the seismicity of the region indicates that the fault system follows the length of the Gulf and enters the Pacific basin south of Baja California. Crustal structure of most of the Gulf is of oceanic type, so that an origin by structural depression of continental rocks is not possible.

Tectonic styles north and south of Los Angeles differ greatly. To the north, the Coast Ranges expose thick Upper Cretaceous and Cenozoic sedimentary rocks that were deposited in local basins and deformed tightly and repeatedly. To the south, in the Peninsular Ranges and Baja California, correlative rocks are thin and show little compressive deformation.

The California batholith of mid-Cretaceous age and allied crystalline rocks form the basement of Baja California, southwestern Arizona, and northwestern Sonora and probably extend along the coast of mainland Mexico; the Gulf apparently bisects the crystalline belt longitudinally.

These features suggest that Baja California initially lay 300 miles to the southeast, against the continental-margin bulge of Jalisco. The Gulf of California may be a pull-apart feature caused by strike-slip displacement plus up to 100 miles of cross-strike separation of the continental plate, subcontinental materials having welled up into the rift gap. The strike-slip motion has a tensional component across the continental margin south of Los Angeles but a compressional component to the north.

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INTRODUCTION

Recent work has demonstrated the probability of a strike-slip displacement of about 350 miles on the right-lateral faults of the San Andreas system; these faults trend longitudinally into the deep and seismically active Gulf of California, which has an oceanic crustal structure, and probably thence into the Pacific Ocean basin. The fault displacement is on the order of the distance between the tip of Baja California and the continental-margin bulge of Jalisco to the southeast. Baja California has the proper shape and, so far as is known, geology to have been part of the mainland, since rifted away by oblique tension across the San Andreas fault system. The striking contrast in tectonic styles of coastal California north and south of Los Angeles can be explained in these terms.

This paper presents the descriptions, based on information published by others, which support this speculation, and notes some mechanical consequences which appear to follow if the explanation is correct. The account advocates as briefly as possible a radical explanation; it does not evaluate at length dissenting arguments. Although the geologic and geophysical data accord remarkably well with this theory, contrary conclusions will be drawn from the same data by many geologists.

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The manuscript was much improved by criticism by Jerry P. Eaton, K. O. Emery, Richard H. Jahns, Philip B. King, and Tj. H. van Andel, and ideas grew during discussions with W. Bradley Myers. Geophysical data provided in advance of publication by J. C. Harrison and George G. Shor, Jr., support the theory advanced here, and Harrison, Shor, van Andel, and others of the Scripps Institution of Oceanography—University of California group have arrived at a similar explanation for the origin of the Gulf of California.

REGIONAL GEOLOGY

The Gulf of California is 700 miles long, outlined by the peninsula of Baja California and the coast of mainland Mexico (Figs. 1, 2); it widens southward to a width of more than 100 miles. Northwest of the head of the Gulf, the structural and topographic province of Baja California continues another 100 miles into southern California as the Peninsular Ranges, and the province of the Gulf continues similarly as the Salton trough; both peninsular and gulf provinces terminate against the Transverse Ranges. The Gulf is shallow at the north end, probably due in part to filling by light sediments, and Upper Tertiary and Quaternary sediments are thick in the Salton trough region. Deep longitudinal and rhombic closed basins within the central part of the Gulf, and the mouth of the Gulf, reach depths greater than 10,000 feet. The common crestal altitude of the peninsula is about 6000 feet, and its highest peaks are higher than 10,000 feet. West of Baja California is a submerged continental borderland with much structural relief in the north but relatively little in the south.

This structural and topographic system—high narrow peninsula, deep narrow gulf, both subparallel to the mainland coast and to the regional continental margin—is strikingly unusual. Beal (1948) concluded that the peninsula is a wedge uplift, bounded by inward-dipping faults on both sides and raised by compression directed normal to the coast; Anderson (1950) expressed the more common opinion that great normal faults separate peninsula and Gulf. However, the vast displacement along the San Andreas fault system has become recognized only recently, and it is essential to incorporate the fault system in any structural considerations. Shepard (1950) did this for the topography within the Gulf, concluding that the topography resulted largely from strike-slip faulting with associated rhombic rifting and also vertical offsets. Carey (1958, Fig. 24) drew the Gulf of California as having originated by the drifting away of Baja California from the mainland, apparently as a silicic raft; he did not suggest that the San Andreas fault might have affected either ocean or mainland outside the Gulf. His suggestion was not appreciably elaborated in the text, appearing chiefly as part of a map illustrating a possible restoration of the Caribbean region according to a jigsaw-puzzle scheme of continental drift. Wegener (1924, p. 184–185) suggested that the Gulf was opened by southward drift of Baja California; perhaps he misunderstood both the location and sense of displacement of the San Andreas fault system.

The active strike-slip San Andreas fault extends most of the length of California before disappearing into the ocean in the north. South of lat 35°N., the San Andreas has major subparallel branches—such as the San Gabriel, San Jacinto, and Elsinore faults—and these, with the San Andreas, lie longitudinally within the California batholith. This great composite batholith of middle Cretaceous age, with other associated crystalline rocks, spans the length of California and Baja California and is probably practically continuous from southern mainland Mexico and the tip of Baja California north through the Idaho batholith. The complexes commonly referred to as the Idaho, Sierra Nevada, Southern California, and Lower California batholiths are only well-exposed portions of a far larger granitic complex.

West of the northern part of this great batholith is the older Nevadan terrane of stocks and small isolated batholiths of granitic rocks, probably mostly of Late Jurassic age, separated by large areas of metamorphic rocks. "Nevadan" has become established in the literature as applying to Late Jurassic orogeny, yet the plutonic rocks of its once-supposed type region, the Sierra Nevada, are for the most part younger (Curtis, Evernden, and Lipson, 1958).

North of the 35th parallel, the San Andreas fault, with most of its offset apparently concentrated in a single narrow zone, separates the Franciscan terrane of unmetamorphosed Jurassic and Cretaceous eugeosynclinal rocks on the east from a slice of the California batholith 350 miles long on the west (Fig. 2). Curtis, Evernden, and Lipson (1958) demonstrated that these granitic rocks of the Coast Ranges west of the San Andreas fault are of the same age as those of the Sierra Nevada. The rocks also are of the same types, dominantly quartz dioritic,

as are those of the western part of the main batholith. Franciscan terrane reappears west of the granitic rocks of the Coast Ranges. As these granitic rocks are younger than most of the unintruded and unmetamorphosed rocks of the Franciscan terrane east of the San Andreas fault, it is highly likely that the post-batholith right-lateral (east-side-south) displacement of the fault is about 350 miles.

Hill and Dibblee (1953, p. 446–449) suggested the following displacements along the San Andreas fault:

(1) middle Cretaceous batholith (then considered Jurassic), 350 miles
(2) upper Eocene sediments, 225 miles
(3) lower Miocene sediments, 175 miles
(4) upper Miocene sediments, 65 miles
(5) Pleistocene gravels, 10 miles
(6) Offsets of late Quaternary drainage lines, ½–1½ miles.

Curtis, Evernden, and Lipson (1958, p. 14–16) presented further evidence for the 350-mile offset of the batholith and mentioned substantiating features of Cretaceous sedimentary rocks. (Because Upper Cretaceous conglomerates on both sides of the San Andreas fault north of San Francisco contain granitic cobbles, Rose [1959] concluded that offset along the fault had not been large; but as granitic source rocks would have been available to both areas prior to any displacement from 0 to 1500 miles, this relationship is irrelevant.)

Hall (1960) found faunal provinces in upper Miocene sediments to be offset at least 50 miles, and possibly 150 miles, by the San Andreas fault. Smith (1959) concluded that Pleistocene movement near San Francisco on the San Andreas fault and a nearby divergent fault totaled at least 15 miles.

Several major faults—San Andreas, San Gabriel, Elsinore, San Jacinto—in southern California share in the aggregate displacement of the San Andreas fault system. Crowell (1960) noted that distinctive Precambrian anorthosite and allied rocks occur, about 150 miles apart, on opposite sides of the fault which carries the name San Andreas, and that a similar offset is shown by pre-Cretaceous schists and by Eocene, Oligocene, and Lower Miocene strata. Crowell (1952) also showed the likelihood of 15–25 miles of post-late Miocene displacement on the San Gabriel fault. Displacements along the Elsinore and San Jacinto faults must be large to account for the great changes across them, but so far as I know specific offsets have not yet been demonstrated. In northwestern Baja

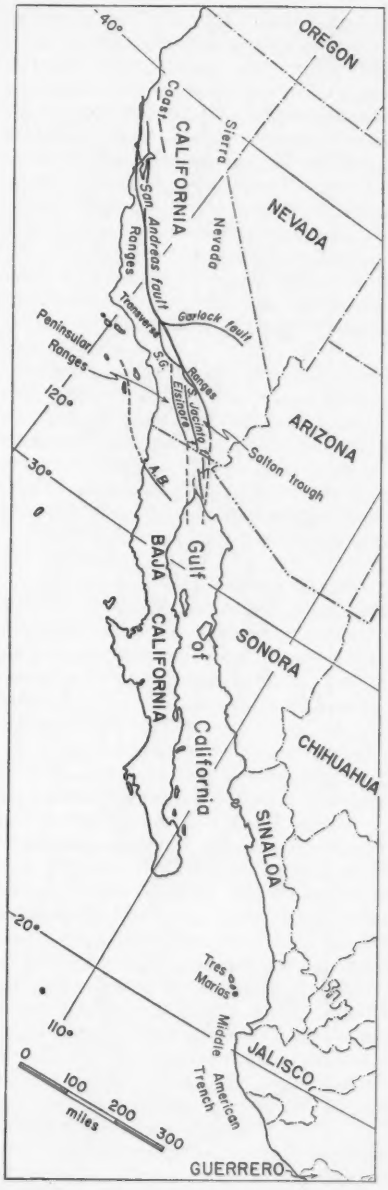

Figure 1. Map of California and Pacific Coast region of Mexico. S. G., San Gabriel fault; A. B., Agua Blanca fault

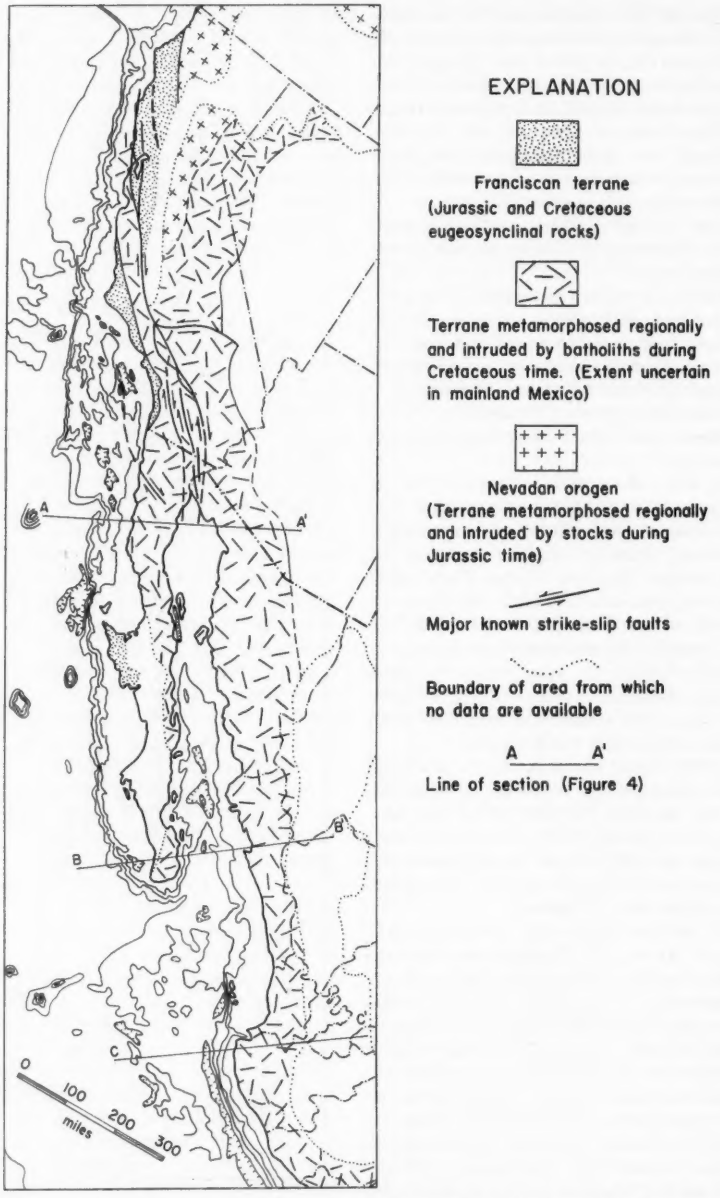

Figure 2. Map of major pre-Tertiary rock complexes of California and Pacific Coast region of Mexico. Submarine contour interval 1000 m

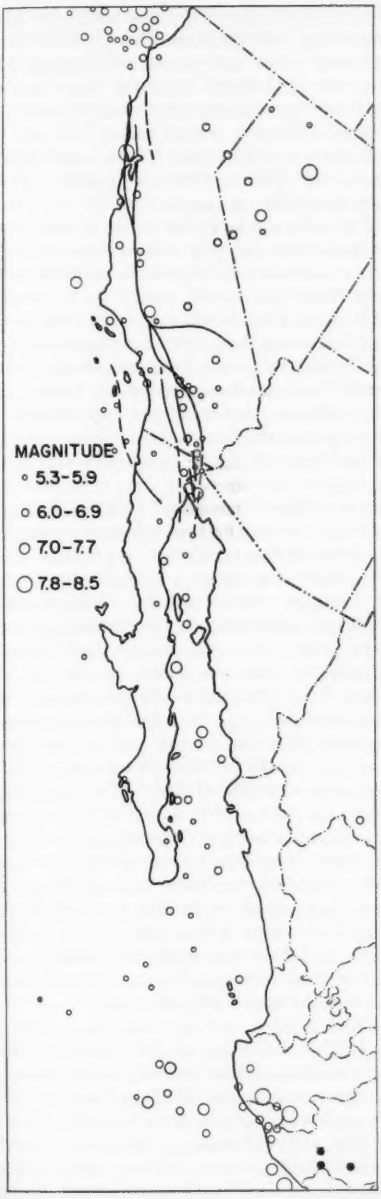

Figure 3. Epicenters of earthquakes in California and Pacific Coast region of Mexico. Open circles, depth shallower than 60 km; solid circles, depth greater than 70 km. After Gutenberg and Richter (1954, Figs. 8, 9)

California, the Agua Blanca fault has a right-lateral displacement that may be greater than 14 miles (Allen, Silver, and Stehli, 1960). Additional strike-slip faulting or major cumulative displacement on rifts subparallel to the San Andreas zone may have occurred throughout the shallow submerged region off California and Baja California (Shepard and Emery, 1941; Emery, 1960, p. 78–79) and the California Coast Ranges. Other geologists, among them J. M. Christie, J. C. Crowell, D. B. McIntyre, and E. L. Winterer, have gathered but not yet published further data indicating the likelihood of great displacements on faults of the San Andreas system.

Strike-slip displacement on the Garlock fault, which trends northeastward from the San Andreas fault, has been left lateral, or opposite in sense to that of the San Andreas system. A distinctive dike swarm is offset 40 miles by the Garlock fault (Smith, 1960), and about the same displacement is indicated by the offset of the east margin of the California batholith (Fig. 2).

Arguments as to the validity of such conclusions will long continue. Woodford (1960) recognized that the right-lateral faults of southwestern California separate very different provinces of pre-Cretaceous rocks, but concluded that the relationships could have been achieved by lateral displacements of only a few tens of miles, combined with huge vertical offsets; such an explanation is inadequate for the sliver of the Cretaceous batholith west of the San Andreas fault in the Coast Ranges.

In another recent paper minimizing displacement, Higgins (1961) stated that post-middle Pliocene offset north of San Francisco had probably been only a few miles. Higgins based this on the assumption that there has been neither appreciable deformation nor erosion since before middle Pliocene time, and that the present landscape is largely a relic from the early Pliocene. His assumptions appear untenable for many reasons: for example, formations younger than the supposed relic landscape have general structural dips of 10°–40° over most of their extent (Weaver, 1949).

Allen (1957) found that the fault previously called the San Andreas west of San Gorgonio Pass in southern California dies out near the pass. Some geologists have misinterpreted his paper as proving that the displacement on the San Andreas system has been minor: Allen (p. 346), however, recognized that he had shown only that a fault had been misnamed and that

this had no bearing on the total offset on faults of the San Andreas system.

The California batholith and the regionally metamorphosed rocks associated with it extend from southern California into Baja California in one direction, and into southern Arizona (Wilson, 1960; Wilson, Moore, and O'Haire, 1960) and Sonora in another. The little geologic mapping done in Mexico suggests that granitic rocks form a semicontinuous belt farther south along the west coast of mainland Mexico but are generally lacking inland, and it is thus likely that the Cretaceous batholithic terrane is continuous from Sonora to central America; but there are few dated rocks in this great expanse, and the metamorphic and granitic complexes have been referred assumptively to various parts of the Precambrian, Paleozoic, and Mesozoic. King (1939, p. 1708) stated that in western Sonora, "The ranges of sedimentary and volcanic rocks appear to be only detached roof pendants in a vast granite batholith or group of coalescing batholiths." The batholith yielded mid-Cretaceous isotopic dates in northwestern Sonora (Giletti and Damon, 1961) and cuts unfossiliferous formations that overlie Lower Jurassic rocks (King, 1939). Anderson (1950) suggested the correlation of the granitic rocks of western Sonora and Sinaloa with those of Baja California. Larsen *et al.* (1958) found the batholithic rocks of Oaxaca and Guerrero, southern Mexico, to be of the same age (mid Cretaceous) as the California batholith.

In California, the batholith is zoned longitudinally, quartz diorite being dominant in the west, leucocratic quartz monzonite in the east. Similarly, the rocks of Baja California are chiefly quartz dioritic (Beal, 1948), whereas those of the mainland coastal belt are chiefly more leucocratic granodiorite and quartz monzonite (Anderson, 1950). It appears that peninsular and mainland granitic rocks, now separated by the deep Gulf of California, are together comparable to the undivided batholith to the north: the Gulf bisects the batholithic belt longitudinally.

The traces of the southern faults of the San Andreas system trend toward the Gulf of California and disappear beneath the delta of the Colorado River. Their continuation can be inferred from seismic data. California earthquakes are concentrated along the San Andreas fault, and Mexican earthquakes are virtually confined to the Gulf of California, throughout the length of which they are common (Fig. 3). Benioff (1959, p. 395) considered the fault zone to be "clearly defined" within the Gulf, and there seems no reason to challenge the generally held belief that the San Andreas fault zone runs the length of the Gulf. Much of Baja California is bounded on the east by fault scarps, and obvious scarps occur within the Gulf (Beal, 1948; Anderson, 1950; Shepard, 1950).

The pattern of Late Cretaceous and Cenozoic sedimentation and deformation in coastal California north of Los Angeles has differed strikingly from that to the south. To the north, sediments accumulated to great thicknesses in local basins and were deformed repeatedly and tightly to produce exceedingly complex structures. The Coast Range mountain blocks, like the dominant local structures, are oblique at low angles to the San Andreas fault and strike farther west of north than does the fault. High-angle reverse faults are extensively developed. The Transverse Ranges, trending westward immediately north of Los Angeles and south of the end of the Coast Ranges, have had a still more complex history and contain, for examples, the remarkable Ventura Basin and broad areas of middle Precambrian plutonic rocks (raised from deep within the California batholith?) uniquely close to the Pacific Ocean; the Transverse Ranges occupy the same latitude range as does the pronounced westward diversion of the San Andreas fault from its generally northwestward course. In the Peninsular Ranges (Jahns, 1954) and Baja California (Beal, 1948), south of Los Angeles, Upper Cretaceous and Cenozoic rocks are thin and have been but gently tilted over large areas, and elsewhere have been deformed by broad and gentle folds and by normal and strike-slip faults. Allen, Silver, and Stehli (1960, p. 461) emphasized "the almost undeformed state of Upper Cretaceous beds along much of the Baja California coast."

These marked contrasts must be related to the Gulf of California, for the southern pattern of thin sedimentation and little deformation is coextensive with the Gulf and its landward continuation. Nor can it be coincidence that the Gulf and peninsular provinces end abruptly at the latitude at which the San Andreas fault swings sharply westward, and that this latitude is one of extreme deformation.

GEOPHYSICAL DATA

Seismic sounding in the axial deep of the Gulf of California, south of about 27°30′ N.,

shows the M discontinuity to be only about 6 km beneath the sea floor, or 8 to 10 km beneath sea level, and the crustal structure to be "very similar to that found in the deep ocean, not withstanding the much shallower depth" (Shor, 1959, p. 2). Sediments are generally thinner than 400 m. In the northern part of the Gulf, the structure is intermediate between those typical of continental and oceanic masses: 5 km of low-velocity sediments overlies 5 km of rocks of sialic velocity ("older sedimentary rock, volcanics or intrusives"), which rests in turn on a layer of intermediate velocity (V_p = 6 3/4 km/sec) (Shor, 1959, p. 2). The M discontinuity is at typical continental depth beneath the Peninsular Ranges, but most of the variation in crustal thickness between there and the continental borderland in one direction and the Gulf of California province in the other is in the intermediate velocity layer (Shor, in press).

J. C. Harrison (Written communication) made gravity measurements and found that the southern part of the Gulf is essentially in isostatic equilibrium: it is deep because it is underlain by heavy rock, and not because it is formed of continental rocks somehow held depressed. The northern part of the Gulf, however, has free-air anomalies which increase in negative magnitude with increasing depth: there is a slight mass deficiency, so that the crust would have to rise somewhat to attain isostatic equilibrium, according to Harrison.

The Imperial Valley, in the Salton trough in the northern extension of the Gulf province, is characterized by a very high heat flow, shown not only by hot springs and related phenomena but also by the hot water encountered in the half dozen or so oil-test wells so far drilled (Robert G. Maynard and Everett W. Pease, oral communication). A very high heat flow also characterizes the deep-water part of the Gulf, according to R. Von Herzen (*in* Shor, 1959, p. 2).

The San Andreas fault system has too great a displacement to die out within the Gulf of California, and available data do not suggest that the system trends onshore anywhere in Mexico. Rather, it appears that the system must strike into the Pacific at the mouth of the Gulf. If the oceanward extension of the San Andreas is marked by earthquakes, then either it continues obliquely seaward through a zone of sparse earthquakes, or it merges with the seismic zone inshore from the Middle American Trench. The similarity of Rayleigh waves, as recorded in Hawaii, generated by earthquakes along the San Andreas fault and within the Gulf of California to those generated in the vague seismic zone trending seaward suggests the seaward continuation of the San Andreas zone, according to Jerry P. Eaton (written communication). On the other hand, the general continuity of trend of submarine scarps in the Gulf of California with the steep, straight scarp which bounds the submarine bank of the Tres Marias islands and defines the northwest end of the Middle American Trench (Fisher, 1961, Plate 2) is better explained by a direct connection of trench and San Andreas structures.

ORIGIN OF THE GULF

The following features bear upon the origin of the Gulf of California:

(1) The southern half of the Gulf has an oceanic crustal structure, and the northern half, including its onshore continuation, has an abnormally thin layer of continental rocks. The Gulf therefore cannot be a structurally depressed block of continental rocks.

(2) West of the Gulf province, Upper Cretaceous and Cenozoic rocks show deformation that is insignificant compared with the effects of strong compression to the north. The change is abrupt and occurs at the latitude of the north end of the Gulf province.

(3) The Gulf province ends to the north against the broad westward deviation of the course of the San Andreas fault, where for nearly 200 miles the fault strikes 30° farther west of north than it does either north or south of that deflection.

(4) Strike-slip displacement on the San Andreas fault system has probably been about 350 miles since mid-Cretaceous time. This displacement, concentrated on a single fault in the north but distributed among several faults in the south, has been achieved progressively over the last 100 million years.

(5) The San Andreas fault system follows the Gulf of California throughout its length and at the mouth of the Gulf probably trends into the Pacific Ocean.

(6) The meager geologic information from Mexico suggests that the Cretaceous batholith of California and allied metamorphic rocks continue along the coast of mainland Mexico. If so, the Gulf of California bisects the crystalline belt longitudinally.

(7) Three hundred miles southeast of the tip of Baja California, in Jalisco, the Mexican

coast swings abruptly southwestward for nearly 100 miles, then as abruptly resumes a straight east-southeast trend. Possibly the southwestward deviation truncates obliquely the west half of the Cretaceous batholithic terrane which reappears, offset 100 miles west of the on-strike position, in Baja California. (Geologic data from Jalisco are needed to evaluate this possibility.)

letting-out of the Baja California block (as, on a strike-slip fault obliquely crossing the peninsula west of the 110th meridian), of thinning of the continental plate, and of sedimentation. Islands and banks in the Gulf and between it and Jalisco may be fragments of the continental plate.

The opening of the Gulf may also be due in substantial part to the oblique orientation of

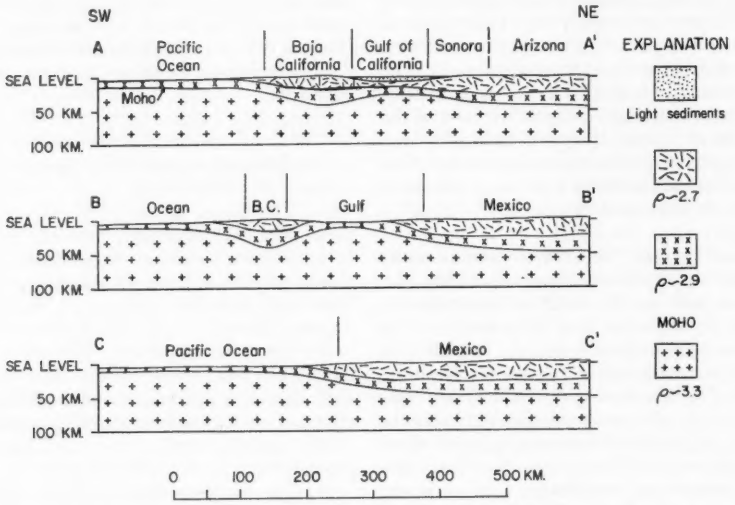

Figure 4. Schematic sections of crustal structure in Gulf of California region. Locations shown on Figure 2. No vertical exaggeration

Each of these features supports the conclusions that Baja California was initially part of the mainland (the tip of the peninsula having been against the bulge offshore from Jalisco) and that displacement along and across the San Andreas fault system caused Baja California to rift from the mainland. If so, then up to 100 miles of cross-strike separation occurred also, and Baja California rotated away from the mainland as it moved north, whereas the two sides of the fault zone remained pressed together in most of California. No contrary theory yet proposed can account for the known features.

The deep-water gap between Jalisco and Baja California is about 250 miles, much less than the 350 miles of displacement on the San Andreas fault in California. The discrepancy may be due to a combination of absorption of displacement by internal deformation, of strike-slip faults in northern Baja California (such as the Agua Blanca fault [Allen, Silver, and Stehli, 1960]) and in southern California and to the westward deflection of the San Andreas fault system at the latitude of the Transverse Ranges.

MECHANISM

Baja California may have formed by the thinning, sundering, and drifting apart of a continental raft of sial, the heavier subcontinental material having flowed into the gap behind the moving plate to form the floor of the Gulf of California. Drift of the peninsula was northwestward, oblique to the continental margin. Over-all rate of the strike-slip component of the motion has been about 1 foot per 60 years (300 or 350 miles in 100 million years), and cross-strike separation rate at the mouth of the Gulf has averaged about 1 foot

per 200 years (total, 100 miles). Present rates may be higher than those of most of the geologic past.

Figure 4 shows schematic crustal sections in the region of the Gulf of California. These are drawn to accord with the geophysical data noted heretofore, with the findings of Press (1960) regarding layering in California and Nevada, with the thin gravity net of Woollard and Monges C. (1956), and with the general empirical relationships between gravity, topography, and depth of M discontinuity inferred by Woollard (1959).

The crustal cross sections are drawn to fit present topography, but they may also represent stages in the evolution of the Gulf of California. The lowest section, across the unbroken continental margin south of the Gulf, indicates the probable predrift structure of the region across the Gulf and Baja California. The uppermost section, near the north end of Gulf and peninsula, depicts an early stage in sundering: the continental raft has been thinned by tensional flow, and the mafic material beneath has flowed upward beneath the thinned zone. The center section shows the extreme stage produced: the continental plate has ruptured, dense rock welling up from beneath has filled the void, and yet the M discontinuity is at the common suboceanic level.

This postulation of upward flowing of mafic material, yet with a resultant M discontinuity at normal oceanic depth, requires that the discontinuity be a zone of density-phase transformation, as of basalt to eclogite, in rock of constant composition in any one column through the discontinuity. This is an old idea, strongly advocated by, for example, Du Toit (1937, p. 230–242), who credited it to papers published in 1913 and 1914 by L. L. Fermor. Geophysical, petrologic, and laboratory data since obtained, and many theoretical arguments of petrogenesis and tectonics, support this interpretation, which is gaining increasing acceptance. A change from mafic to ultramafic rocks perhaps occurs at a depth near 400–600 km, where major discontinuities have been suggested by some geophysicists.

Another requirement of the drift theory presented here is that simatic rocks must be able to deform by flow at depths perhaps as shallow as 10 or 15 km.

A qualitative mechanical model can be imagined if we assume that the strike-slip faulting is caused by transport above flow currents in deeper materials. Velocity would increase downward as drag effects reduced surface velocities substantially below those at depth. Velocity of flow must change also across the strike of the fault system, for strike-slip faulting at the surface is accomplished along a number of subparallel faults. Velocity thus should increase obliquely downward. Figure 5 illustrates this schematically and is based upon the assumptions that the ocean block is moving, that the rate of increase of velocity is greater laterally than vertically at shallow depths but equal to it at deeper depths, and that the subcrustal flow is laminar. As layers of laminar flow are parallel to velocity contours, or perpendicular to the direction of increasing velocity, they would be inclined in the sense shown. Different assumptions as to relative magnitudes of vertical and horizontal velocity gradients would of course change the slope of the flow planes; were the continental block the moving one, the direction of slope would be reversed.

The direction of subcrustal flow must vary relative to the strike of strike-slip faults at the surface. North of Los Angeles, the tight compressional structures, oblique at low angles to the strike-slip faults, and the lack of separation of the continental plate, show that the flow has a compressional component directed toward the faults. South of Los Angeles, the tensional structures and the apparent continental rifting indicate that subcrustal flow has a tensional component away from the continent. The theory for the origin of the Gulf by rifting thus explains also the striking contrast in tectonic styles of the coastal regions north and south of Los Angeles. The extreme deformation of the Transverse Ranges occurs where the compressive component of flow is greatest—where the northward-drifting block of Baja California strikes the westward deflection of the San Andreas fault system.

Regardless of the validity of this theory, and regardless of whether the San Andreas fault system has had 350 miles of displacement, as accepted here, or only the few tens of miles of offset that many conservative geologists are willing to accept, it is obvious that the Cenazoic compressive structures of the California Coast Ranges are by-products of strike-slip faulting, caused by compression oblique at a low angle to the strike-slip faults, rather than by compression at a high angle as is often assumed.

Right-lateral strike-slip faulting has been recognized in a number of places about the

Pacific Ocean, and a number of geologists and geophysicists (as for example, Benioff, 1959) have suggested that the ocean basin is rotating counterclockwise relative to the continents. Earthquake first-motion studies suggest a dominance of strike-slip movements along the

Oblique tensional rifting may be a common process. Left-lateral strike-slip faults trend from the Dead Sea trough into the long and narrow Red Sea, which has an oceanic structure like that of the Gulf of California; if the origin is similar, then Africa is moving relatively

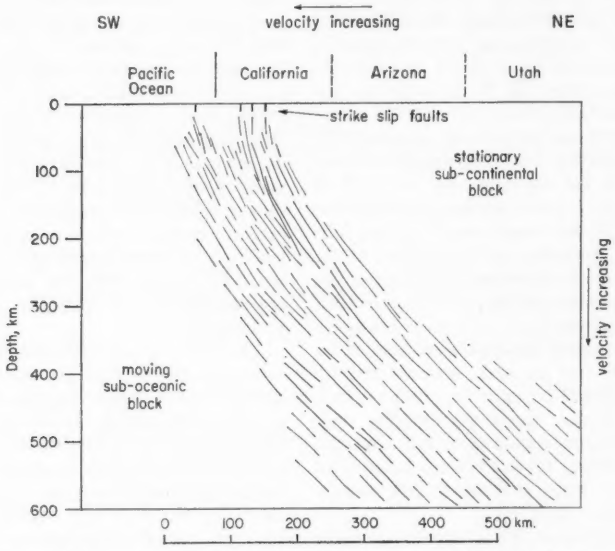

Figure 5. Cross section showing possible orientation of planes of laminar flow beneath the San Andreas fault system. Motion largely strike slip, perpendicular to the plane of the section

structures which dip beneath the island arcs and the continental margins along the ocean trenches. The mechanism of San Andreas faulting inferred in Figure 5 could produce both ocean trenches and continental strike-slip fault systems as parts of a continuous movement system. Continuity between trenches and strike-slip faults can be inferred in other places about the Pacific: for example, the Alpine fault of New Zealand, which has a right-lateral displacement comparable to that of the San Andreas fault, probably trends into theTonga Trench.

southward away from Asia Minor and the Arabian Peninsula. The Cayman Deep in the West Indies may be another tensional rift, and the Snake River depression of southern Idaho is perhaps a rift still in an early stage. Many other examples could be cited. The possibility of great lateral transport and rotation of continents and ocean basins, and of any portions of each, adds disconcerting variables to structural interpretations but provides a way to resolve the increasingly perplexing problems which are insoluble with the conventional assumption that lateral shift is of little consequence.

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Arctic Basin Geomorphology

Abstract: The SSN NAUTILUS obtained the first continuously recorded echo-sounder profile across the center of the Arctic Basin during its 1958 cruise from Point Barrow to the North Pole and then to the Atlantic Ocean. The sounding lines, supplemented by lines from SSN SKATE reveal the following geomorphic features: (1) the Alaskan continental terrace and continental rise; (2) Canada Basin, a flat abyssal plain at the depth of 3850 m; (3) Chukchi Cap, a projection from the Chukchi Shelf with a summit elevation of about 900 m; (4) a continuation of Canada Basin, interrupted by a few sea knolls and having a depth of 3900 m; (5) Central Arctic Rise, a region of smoothly undulatory topography at depths between 2000 and 3700 m, with relief not exceeding 900 m; (6) Central Arctic Basin, an abyssal plain at 3940 m; (7) Lomonosov Ridge, rising abruptly from the basin to a summit at 1290 m; (8) beginning almost at the North Pole, the Eurasia Basin, an abyssal plain sloping south with a depth gradually increasing from 4090 m to 4500 m; (9) a region of rough topography containing numerous sharp peaks and having relief of as much as 1100 m and containing the deepest sounding of the Arctic Basin, 5335 m; (10) Nansen Ridge, a smoothly undulatory ridge rising to 1280 m where crossed.

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INTRODUCTION

In August 1958, SSN (571) NAUTILUS traversed the Arctic Ocean beneath the permanent ice pack by diving north of Alaska and surfacing nearly 3700 km away in the vicinity of Spitzbergen, passing the North Pole en route. Along with other scientific measurements, the NAUTILUS obtained an almost continuous echogram which provided the first continuous bathymetric profile across the Arctic Ocean[1]. This profile is oriented favorably for geomorphic reconnaissance, for it transects almost perpendicularly the major ridge and the major rise which traverse the Arctic Ocean Basin and crosses the three secondary basins within the Arctic Ocean Basin. SSN (578) SKATE also made an extensive foray under the permanent ice pack in August 1958, stopping at the U.S. IGY ice drifting station Alpha and reaching the North Pole also. SKATE made a second under-ice journey in March 1959. SKATE's soundings were taken by a Precision Depth Recorder; hence, the echograms are particularly valuable for geomorphic study.

By means of abundant isolated spot sound-

[1]The bathymetric data are published here in accordance with U. S. Navy OPNAV Instruction 3160.6A dated 20 January, 1960.

Geological Society of America Bulletin, v. 72, p. 1319–1330, 4 figs., 4 pls., September 1961

ings, Soviet workers have outlined in a rough way the major geomorphic features lying beneath the Arctic Ocean (Fig. 1). Workers on ships frozen in the ice pack, on ice island drifting stations, and on icebreakers on the periphery of the permanent ice pack have studied the sea floor in limited regions of the Arctic Ocean in greater detail. The development of the nuclear submarine has at last provided a really suitable method for obtaining bathymetric data comparable with those that can be obtained by surface ships in other seas of the world. Anderson (1959) and Lalor (1959) give popular accounts of nuclear submarine journeys.

This paper was presented at the First International Symposium on Arctic Geology, held at Calgary, Alberta, Canada, on 11–13 January, 1960. The University of Toronto Press is publishing a volume including most of the papers presented at the symposium.

ACKNOWLEDGMENTS

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NOMENCLATURE OF BATHYMETRIC FEATURES

At present, owing to the lack of exchange of information among Canadian, Soviet, and United States personnel, there is little agreement on names for the prominent bathymetric features of the Arctic Ocean. Dr. F. G. Markov, of the Institute of the Geology of the Arctic in Leningrad, informed Shumway (Personal communication) that Soviet workers have named these features, but workers in the United States and Canada are unaware of what these names are, except for the well-publicized Lomonosov Ridge. We have received no reply to a letter of inquiry concerning these names that was sent to this Soviet institute in February 1960. We feel that, while precedence in the use of names is important, occasions may arise in the early stages of discovery and description of bathymetric features when it is best to give new names with preference of geographic names for the sake of clarity, consistency, and brevity.

HISTORY OF ARCTIC BASIN SOUNDINGS

Fridtjof Nansen discovered the Arctic Basin in the scientific sense, for he first learned that deep water existed beneath the polar sea ice. This apparently came as a surprise, for he was prepared to sound only to 500 m. Under Nansen, the FRAM, during her 3-year drift (1893–1896) in the pack ice, recorded eight wire soundings which attained bottom between the New Siberian Islands and the Greenland Sea (Nansen, 1904). In 1908, Perry attempted a deep cast near the North Pole, but his line broke at 2750 m. In 1913, nine soundings were collected from the KARLUK in open water north of Alaska before she became imprisoned and finally crushed by the ice. From 1935 to 1939 the Soviet ice breakers G. SEDOV and SADKO obtained about 55 soundings off the Siberian arctic sector. Both ships became imprisoned in the ice pack. The SADKO was eventually free, but the G. SEDOV drifted across the polar basin to Greenland.

Wilkins (1928) initiated a new technique when he landed an airplane 550 miles northwest of Point Barrow on the pack ice. His single sounding of 5455 m, obtained with a crude echo sounder, is the deepest claimed for the Arctic Basin. However, shallower soundings obtained nearby by the Soviet Ice Drifting Expedition (North Pole #2) indicate that Wilkins' sounding is erroneous. An extension of Wilkins technique was employed by four Soviet scientists under Papanin (1947), who in 1937–1938 landed on an ice floe near the North Pole and then drifted down along Greenland. Sixteen deep soundings were obtained by Papanin's North Pole #1 expedition. Emery (1949) published a chart of the Arctic Basin based on 152 soundings deeper than 915 m, or 1 per each 12,000 square miles, which summarized our knowledge of the polar basin at that time.

ORGANIC PRODUCTION IN THE ARCTIC

The Precision Depth Recorder, by insonifying a large cone of water beneath the submarine, gives an integrated picture of the concentration of zooplankton. Throughout the under-ice cruises, no discrete hyperbolic echoes were recorded such as are commonly found in ice-free oceans and which indicate schools or

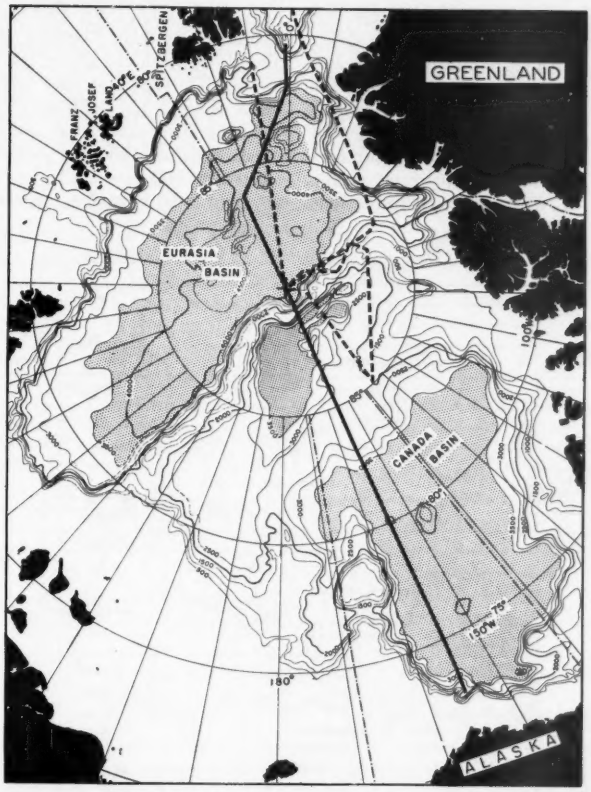


Figure 1. Tracks of SSN (571) NAUTILUS (solid line) and SSN (578) SKATE (dashed line) in the Arctic Ocean during August 1958. Bathymetry from published Soviet soundings (Canada Defense Research Board, 1957)

individual fish, squid, or other nekton. Nor were diffuse clouds of scatterers recorded, such as constitute the deep scattering layers of microplankton which are well known from other oceans.

This evidence suggests that the deeper waters of the Arctic Ocean are among the most sterile in the world and have a lower organic production than even the most lifeless regions of the Central South Pacific. Fishes such as the arctic cod are present in the near-surface water, however, and are the main food supply for the arctic seals which in turn support the polar bear and the arctic fox. Some small fish were seen from time to time in the externally mounted closed circuit television system and through the periscope.

NAUTILUS AND SKATE ECHO-SOUNDING PROFILES

The track of NAUTILUS across the Arctic Ocean begins almost directly north of Point Barrow, Alaska, and runs in a straight line past the North Pole to about 86° north latitude and then turns to the right to end over the Nansen Ridge between Greenland and Spitzbergen (Fig. 1). Superposition of this track on a bathymetric chart of the Arctic Ocean prepared in Canada from published Soviet soundings (Canada Defense Research Board, 1957) shows that this track crosses the major bathymetric features of the Arctic Ocean (Fig. 1).

The NAUTILUS was equipped with an

AN/UQN 1 echo sounder. Unfortunately this was operated on the 6000-fm scale (*i.e.*, about 33 cm of paper width represents 6000 fm), and therefore the echograms are not suitable for direct reproduction. Also, because the submarine cruised at varying depths, echogram depths must be corrected for the depth of submergence. Figure 2 shows a bathymetric profile along the track of NAUTILUS with corrections for submergence, and with the assumption of a sounding velocity of 4800 ft/sec.

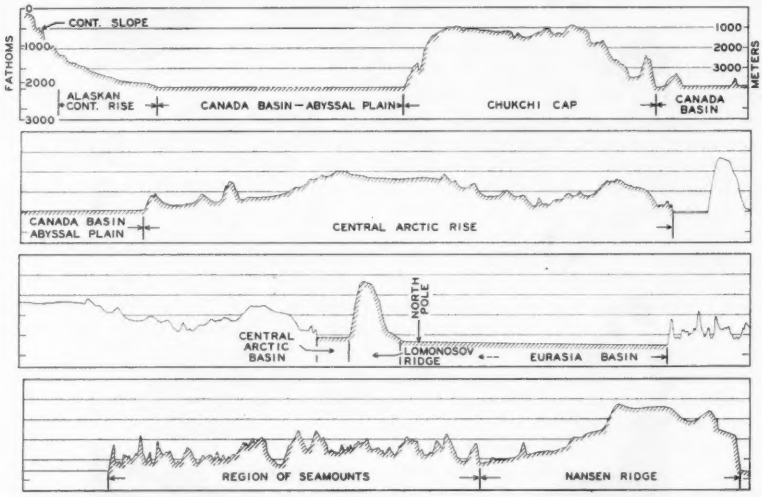


Figure 2. Bathymetric profile across the Arctic Ocean based upon soundings from SSN (571) NAUTILUS. Vertical exaggeration approximately 26 ×. For location of track, *see* Figure 1.

The track made by the SKATE in August 1958 lay almost entirely between the NAUTILUS track and the Canadian archipelago and Greenland (Fig. 1). SKATE entered and left the Arctic Ocean through the Greenland Sea. She visited U. S.-IGY ice drifting station Alpha and did considerable sounding in the vicinity of Lomonosov Ridge with a Precision Depth Recorder. Portions of SKATE's echograms are presented in this paper to illustrate certain geomorphic features (Pls. 1–4).

GEOMORPHIC PROVINCES

CONTINENTAL SLOPE AND RISE NORTH OF ALASKA: NAUTILUS entered the Arctic Ocean through the Barrow Sea valley, which provides a relatively deep avenue of access to the deep ocean north of the Chukchi Shelf. According to A. J. Carsola (1953, Ph.D. thesis, Univ. of California at Los Angeles) this valley lies directly beneath the axis of the North Alaskan Littoral Current and may have formed as a result of scour and differential deposition.

The continental slope and rise north of Alaska constitute the first geomorphic province crossed by NAUTILUS. The rise abuts the slope at about 1800 m and extends with a concave upward profile for about 100 km, where it meets the flat floor of Canada Basin at a depth of about 3810 m (Fig. 2). Fisher, Carsola, and Shumway (1958) have also published three profiles across this continental rise and slope and a contoured chart of a portion of it, on the basis of work done from icebreakers. Carsola *et al.* (1961) present four additional profiles and a second contour chart, also made from icebreaker data.

The profile of this slope and rise is roughly similar to that along the east coast of the United States. Presumably this rise is a great apron of detritus shed from the continent. According to Dietz (1952) such detrital aprons are characteristic of continental terraces in their mature stage of development. Probably a similar apron surrounds much of the Great Arctic Basin. The writers know of no evidence suggesting trenches around the edge of the basin.

CANADA BASIN: The second province crossed by NAUTILUS was the abyssal plain of Canada Basin, which lies at a depth of about 3850 m (Fig. 3). This abyssal plain extends northward for about 460 km along NAUTILUS' track to a massif apparently connected to the Chukchi Shelf, here called Chukchi Cap (at the suggestion of B. C. Heezen). The abyssal plain of Canada Basin extends for an additional 380 km beyond Chukchi Cap. Published soundings show that the two abyssal plains are parts of the same basin and that Chukchi Cap extends into the basin. NAUTILUS' echogram indicated an identity of depth for the two portions of the abyssal plain. The bottom of the Canada Basin extends about 1100 km north-south.

The abyssal plain of Canada Basin has a strikingly smooth and flat sea floor. Fisher, Carsola, and Shumway (1958) obtained soundings on icebreakers for about 100 km into the basin and commented upon its flatness. The southern portion of the NAUTILUS profile shows no topographic interruptions; the northern portion reveals only two sea knolls. This attests to a thick sedimentary fill which has smothered all minor initial irregularities. The flatness is so striking that a single sounding of 3820 m is sufficient to define the depth of the basin.

Figure 3. Geomorphic provinces of the Arctic Ocean Basin

Fisher, Carsola, and Shumway (1958) refer to this basin as Beaufort Basin, and the U. S. Navy Hydrographic Office Charts 6767 and 6768 refer to it as North Canadian Basin: however we feel that the name Canada Basin seems most appropriate.

CHUKCHI CAP: The large massif which interrupts the NAUTILUS track across Canada Basin has a summit level at a depth of about 1100 m

and a minimum depth of 860 m. It rises abruptly from the floor of the basin and has a somewhat rounded and undulatory top. The higher parts of the summit suggest truncation, implying that possibly the massif was subjected to surf planation at some earlier period of *relatively* lower sea level. However, this is as yet only a speculation.

Chukchi Cap, as an elevation of the sea floor, is connected with the Chukchi Shelf. The bathymetric contours shown in Figure 1 indicate the presence of a saddle between the shelf and the cap. On some U.S. Navy Hydrographic Office charts the seaward projection of Chukchi Shelf and the Cap are together called Chukchi Rise. The information now at hand indicates that the term rise is inappropriate.

The 3000-m-high slopes of Chukchi Cap have the appearance of continental slopes. This and the relatively shallow summit elevation suggest that the massif is an outlier of the continental shelf and probably of similar rock structure. Northwind Seahigh (Fisher, Carsola, and Shumway, 1958), a smaller prominent elevation within Canada Basin but connected to Chukchi Shelf through a saddle, is present southeast of Chukchi Cap. This feature, although smaller than Chukchi Cap, has similar

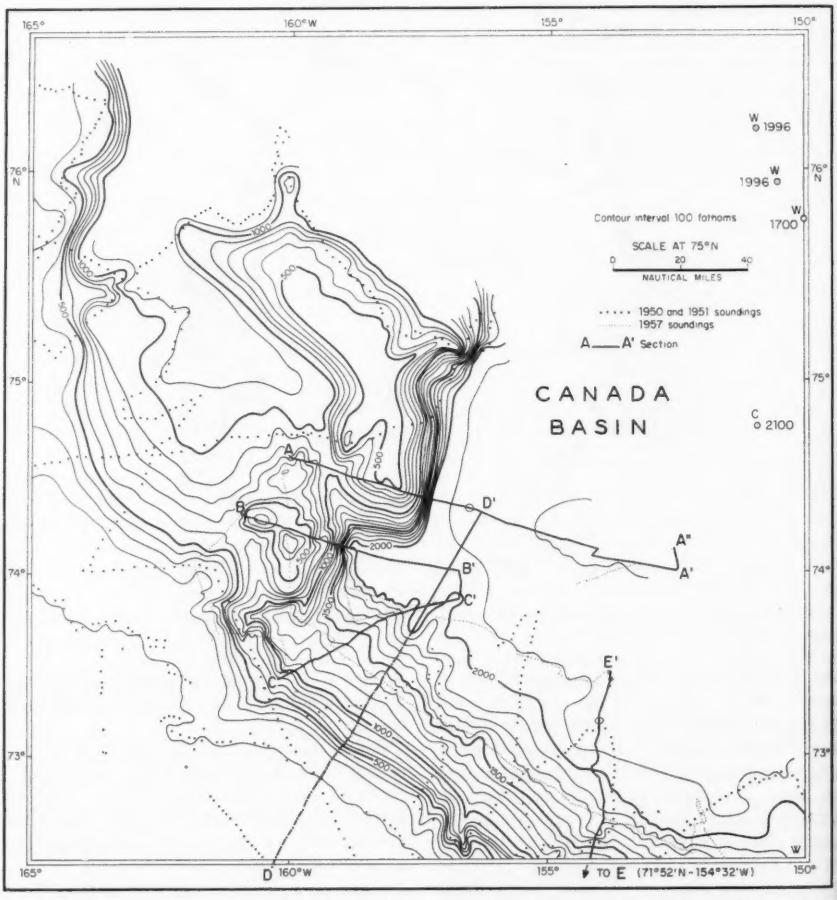


Figure 4. Bathymetric contours of Northwind Seahigh, a feature near Chukchi Cap (From Fisher, Carsola, and Shumway, 1958, p. 3)

steep sides, summit elevations, and surface relief (Fig. 4). Taken together, the two elevations constitute a region of intermediate depth adjacent to a continental shelf. Shepard (1948, p. 105) has applied the term "continental borderland" to similar regions. Although the term leaves something to be desired, inasmuch as "land" is not involved, the geomorphic concept of a sea-floor area of intermediate depth adjacent to a continental shelf, *that has essentially continental-type crustal structure* seems useful. The California Continental Borderland and Blake Plateau are the best known examples.

CENTRAL ARCTIC RISE: The fourth province NAUTILUS crossed is a broad, smoothly undulatory rise about 900 km across and with a summit at about 1800 m. Other data (*e.g.*, Fig. 1; International Geophysical Year, Bull. 6, 1957; Bull. 22, 1959) show that this rise extends entirely across the Arctic Basin. The rise does not merge smoothly into the abyssal plains at either flank, but rather is abruptly separated by escarpments about 600 m high which presumably mark major faults. On an average the rise is about 900 m shallower than the basins on either side. The U.S.-IGY drifting station on ice island T-3 passed over a portion of the rise near Ellesmere Island between longitudes 80° W. and 100° W. (Crary, 1954; Crary and Goldstein, 1957). The SKATE crossed this rise twice (Fig. 1). The U.S.-IGY Ice Drifting Station Alpha spent much of its time over the rise, and a number of geological and geophysical studies were made (International Geophysical Year, 1957; Hunkins, 1961; Schwarzacher and Hunkins, 1961).

The middle part of the Central Arctic Rise, where crossed by the Nautilus track, is a high plateau about 100 km across with a general depth of about 2300 m and with one region reaching to about 2000 m depth (Fig. 2). On either side of this plateau the sea floor drops off gradually to the abyssal plains of Canada Basin and Central Arctic Basin, but the flanks of the rise contain relatively large undulations, with relief of as much as 1000 m. On the north side of the rise a particularly large elevation is present just before the Central Arctic Basin is reached. Possibly this is an elongate feature paralleling the rise and having a similar trend. Notably absent from the rise is jagged topography, usually associated with sea-floor vulcanism.

A portion of a SKATE echogram across the rise shows in detail the top of one of the higher elevations (Pl. 1). The rather flat top of this feature lies at about 1340 m; the highest point reaches to a depth of about 1250 m. The flat-toppedness of this feature suggests that possibly it was wave-cut at some past time.

Little can be said with certainty about the origin of this rise other than to suggest that vulcanism has played no important part in its construction. The subdued and smooth topography of the knolls on the rise as well as the lack of any known seismic activity suggest that the structure is not a modern one. The rise may have been formed by a broad uplift of the sea floor, limited on both sides by major fracture zones along which considerable vertical displacement has taken place.

CENTRAL ARCTIC BASIN: The fifth province crossed was the abyssal plain of the Central Arctic Basin which has an extent along the NAUTILUS track of only 65 km between the Central Arctic Rise and Lomonosov Ridge. Where traversed, the basin floor is completely featureless and flat at 3940 m, or 120 m deeper than Canada Basin. Other existing sounding data show that the Central Arctic Basin is by far the smallest of the three basins beneath the polar sea. Toward Siberia it apparently broadens into a considerably wider basin. The NAUTILUS track traversed only the neck of a much larger pear-shaped basin. Doubtless, the smooth floor of the basin is due to the entrapment of a considerable thickness of sediment.

The nature of the flat floor of this basin and of the way it joins the steep scarp of Lomonosov Ridge is illustrated by a portion of a SKATE echogram (Pl. 2).

LOMONOSOV RIDGE: Shortly before reaching the North Pole, SSN NAUTILUS crossed Lomonosov Ridge. The position and depth recorded agree well with Soviet data (Fig. 1). Soviet literature (*e.g.*, Burkhanov, 1957) states that this ridge was discovered in 1948 and subsequently named Lomonosov Ridge after the great 18th Century Russian physicist, grammarian, and founder of Arctic oceanography. The Soviets have shown that this ridge extends from one side to the other of the Arctic Ocean Basin, commencing off the New Siberian Islands and extending to the shelf off Ellesmere Island.

Lomonosov Ridge rises abruptly from the abyssal plain of the Central Arctic Basin (Fig. 2; Pl. 2) to a rounded summit at a depth of 1290 m where the NAUTILUS track crossed it. Thus the profile has a relief of more than 3 km at this location. The south-facing flank of the ridge is a scarp with a declivity of 13°. The

north-facing flank is less steep, with the region shallower than 3850 m being steeper than the region from that depth to the abyssal plain floor of Eurasia Basin which lies at about 4300 m.

The flanks of the ridge are slightly convex upward, which rules out the possibility of the ridge having been constructed by volcanic processes. The asymmetry of the ridge, the slope convexity, and the overall topographic simplicity suggest that the feature is a great fault block or a slightly overturned and folded geanticline.

The NAUTILUS track shows a summit about 26 km wide which appears to be remarkably flat and suggests that the top of the ridge was truncated at some earlier time by surf cutting when sea level was nearly 1400 m relatively lower than at present. This truncated aspect of Lomonosov Ridge is amply confirmed by SKATE echograms (Pl. 3). The shallowest depth along the central part of the ridge is 936 m according to Soviet sources up to 1954. The minimum depth found by SKATE is 1090 m, in the region near the North Pole. A sounding of only 504 m was found by SKATE, but this was close to the continental shelf on the Canadian side. The depth of the ridge summit seems to be fairly uniform between 900 m and 1450 m, and the maximum sill depth is about 1650 m.

Secondary protuberances on the summit of the ridge such as might be formed by volcanoes are remarkably rare. This further adds to the probability that the summit has been planed off by surf action. Hakkel (1958) discussed evidence for recent submarine volcanic activity on Lomonosov Ridge and suggested that a volcano exists at a depth of 1463 m at 88°16′ N., 64°35′ W. His evidence, however, is tenuous. He ascribes a strong shock felt on an ice floe to submarine vulcanism when ice floe movement or breakup could be the explanation.

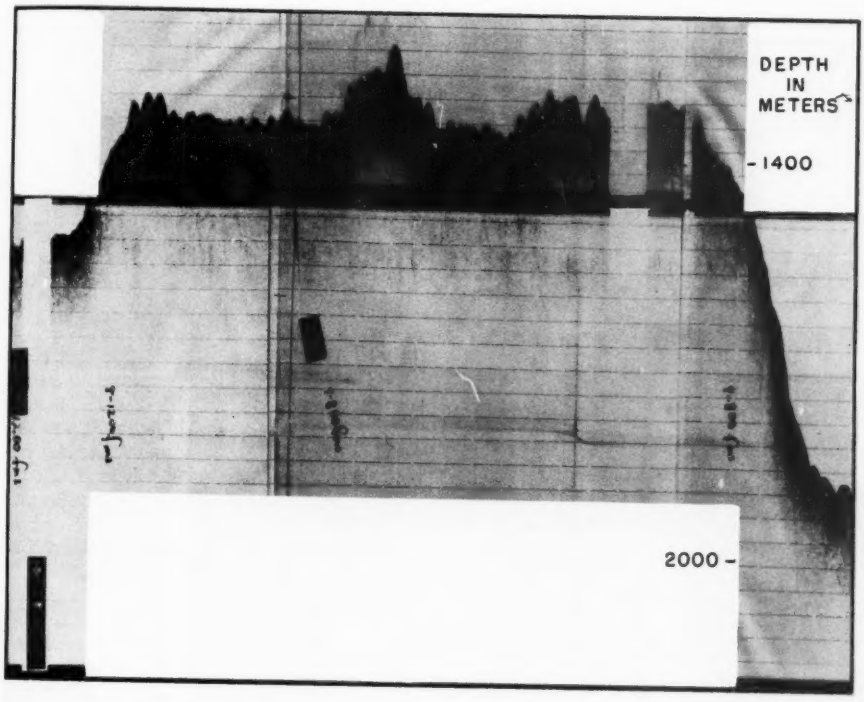
Unaware of earlier Soviet soundings, Worthington (1953) predicted the existence of a ridge extending from Ellesmere Island to the New Siberian Islands on the basis of the 0.50° C warmer water temperature of the deep water of the Alaska side of the Arctic Ocean compared with that on the European side. He further predicted that the maximum depth (the sill depth) of this barrier to free circulation should not be more than 2300 m deep. This agrees reasonably well with the bathymetry as it now is known.

The Lomonosov Ridge structure probably can be traced through both the New Siberian Islands on the Soviet side and Ellesmere Island on the Canadian side. Saks *et al.* (1955) assigned a Mesozoic age to Lomonosov Ridge, pointing out that the New Siberian Islands orogeny is Mesozoic. The folded structures on Ellesmere Island are said to be Mesozoic, too, but according to Eardley (1951) the date of this orogeny is not clearly established. On the continents, such folded mountain structures generally are eroded down to roots even as they rise, but beneath the sea erosive degradation is ineffective. Lomonosov Ridge, then, uniquely may be a great fold in the earth's crust preserved in almost pristine condition. As such, it deserves very detailed geological and geophysical exploration.

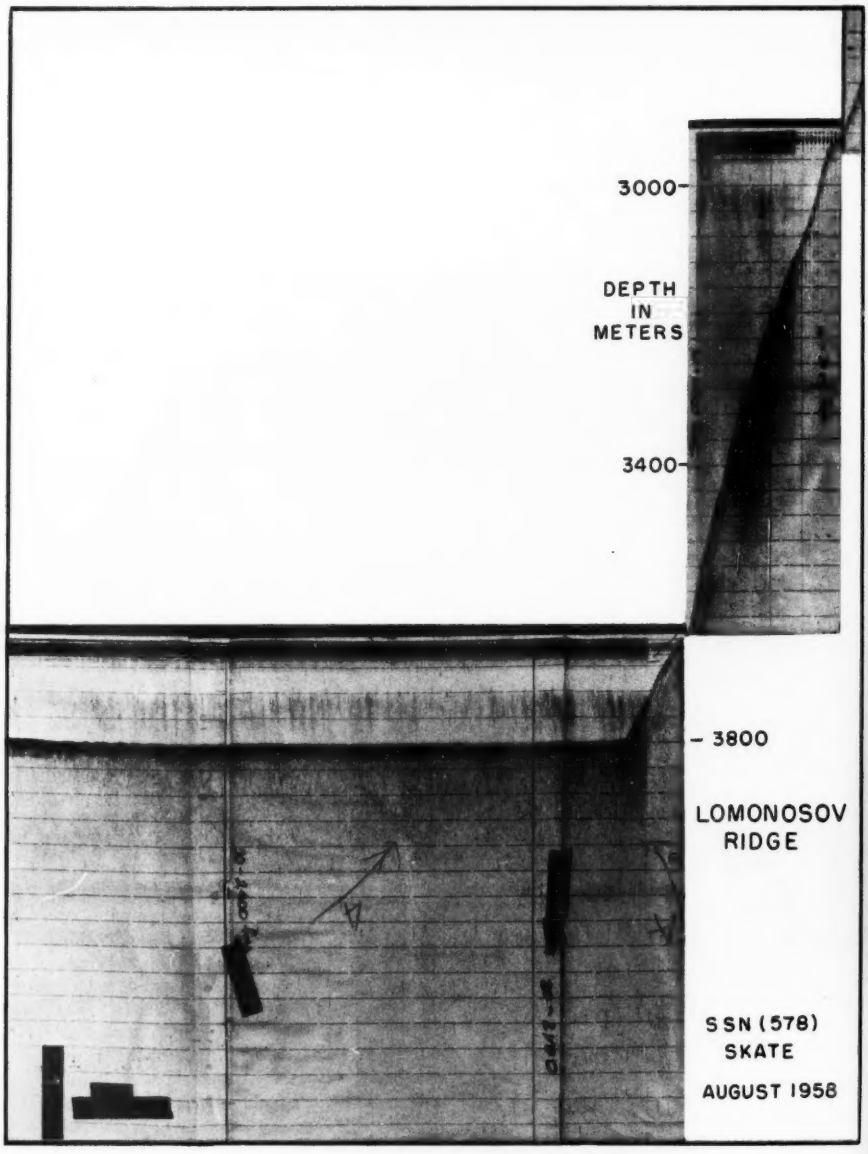
EURASIA BASIN: On the Eurasian side of Lomonosov Ridge lies the abyssal plain of Eurasia Basin (called North Eurasian Basin on some U.S. Hydrographic Office charts). This basin is about equal in size to Canada Basin, but NAUTILUS' track ran along it for only 480 km because the course was approximately normal to the major axis of the basin (Fig. 1). As with the other abyssal plains of the Arctic Ocean Basin, its most striking aspect is its flatness and lack of relief, which indicates a considerable thickness of sedimentary fill. A portion of a SKATE echogram showing the nature of the abyssal plain floor in the region of the North Pole is shown in Figure 1 of Plate 4. The apparent change in depth of the bottom near the North Pole is due to a change in the submergence depth of the submarine. The apparent minor undulations of the sea floor result from minor fluctuations of the submergence depth.

The North Pole is located close to the juncture of Eurasia Basin and Lomonosov Ridge. At the Pole NAUTILUS measured the depth to be 4290 m, but some inaccuracies were involved. After corrections for temperature, pressure, and salinity to a SKATE echogram, the depth is 4200 m, which probably is an accurate value. In 1952 an airplane associated with Ice Island T-3 landed at the North Pole and obtained a sounding of 4300 m (Crary, 1954). Nearby, at 88°54′ N. and 20° W. a sounding of 4290 m was made during the Soviet NORTH POLE drift (Papanin, 1947).

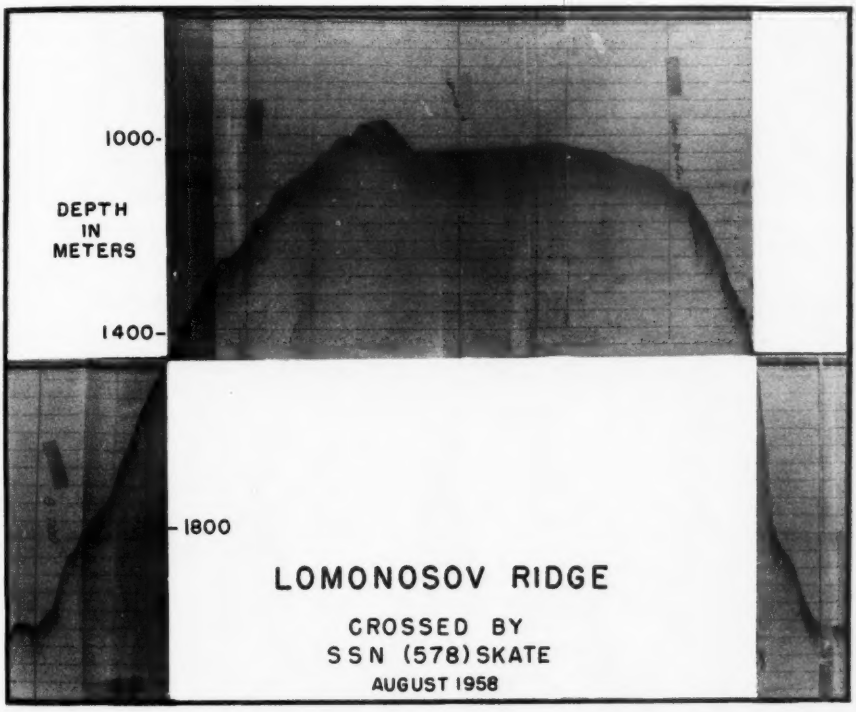
In the vicinity of the North Pole, the abyssal plain is flat as well as smooth, but a bit farther south it dips southward, with the depth increasing gradually from 4090 m to 4500 m, an apparent depth change of about 0.5 m per km. It is, of course, not known at what angle the



PORTION OF SKATE ECHOGRAM TAKEN ON CENTRAL ARCTIC RISE
Average depth along the rather flat top about 1340 m



PORTION OF ABYSSAL PLAIN OF CENTRAL ARCTIC BASIN AND STEEP FLANK OF LOMONOSOV RIDGE AS SHOWN BY PART OF AN ECHOGRAM FROM SSN SKATE



BATHYMETRIC PROFILE ACROSS SUMMIT OF LOMONOSOV RIDGE
AS REVEALED BY A PORTION OF AN ECHOGRAM FROM SSN SKATE

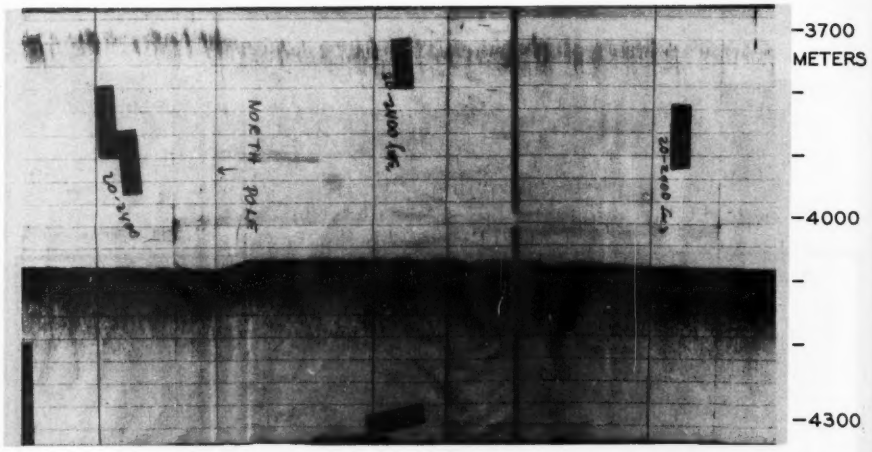


Figure 1. Portion of SKATE echogram taken over Eurasia Basin. Apparent sea-floor-depth change near North Pole and minor undulations result from changes in submarine's submergence depth.

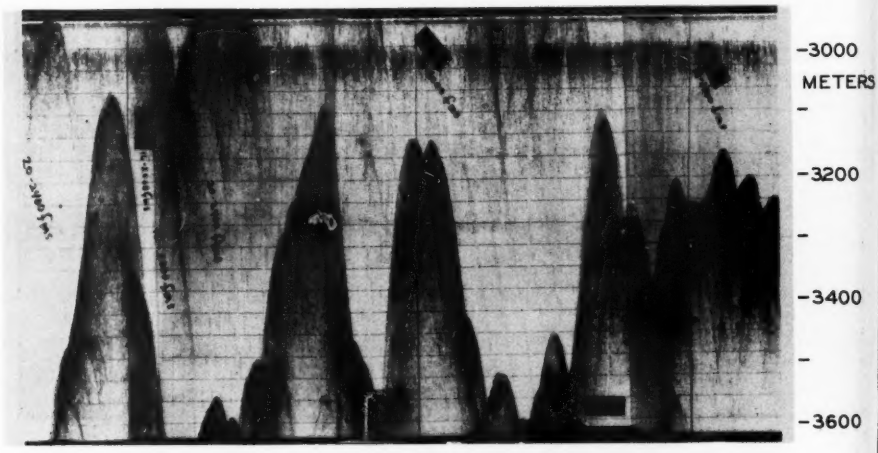


Figure 2. Portion of SKATE echogram taken in region of rough topography north of Nansen Ridge, in Arctic Ocean

PORTIONS OF SKATE ECHOGRAMS TAKEN OVER EURASIA BASIN (VICINITY OF NORTH POLE) AND NORTH OF NANSEN RIDGE (IN ARCTIC OCEAN)

NAUTILUS track lies with respect to the maximum dip. With the facts at hand it is not possible to explain the slope. Small dips of a sedimentary-basin floor may in some cases be related to the sediment-source direction. In this case Lomonosov Ridge lies up-dip, and it does not appear to be a major sediment source. Possibly the dip is unrelated to sediment-source direction and instead is due to tectonic subsidence. The slope is inclined toward the Region of Seamounts. These seamounts are youthful and probably in part volcanic. The extrusion of lava tends to leave the crust unsupported, and the building of volcanoes impresses an added load on the crust. The crust may then tend to sink regionally over an area extending far beyond the confines of the seamount province. The situation is somewhat reminiscent of the crustal failure extending out from the Hawaiian Ridge.

REGION OF SEAMOUNTS: The flat, slightly tilted floor of Eurasia Basin ends abruptly at the beginning of a province of jagged topography (Fig. 2) containing a continuous string of peaks of various sizes and having a maximum relief of about 1000 m. For purposes of convenience, we refer to this province as the Region of Seamounts, although it is not known whether the peaks are conical seamounts or cross sections of ridges. A more formal name for the region should await better data.

Comparison with the topography of the Central Arctic Rise make certain differences striking. First, the peaks in the Region of Seamounts are not superimposed on an over-all rise or swell of the sea floor; instead they rise more or less individually from a general level of about 4200 m, about the depth of Eurasia Basin. Second, the topography of this province has greater amplitude and roughness. The SKATE recorded a sounding of 5335 m in the Region of Seamounts, which is the deepest sounding in the Arctic Ocean Basin confirmed by modern echo sounders to our knowledge.

The topographic roughness suggests vulcanism, but many of the peaks are asymmetric, which suggests a fault-block origin. The province can be described as a youthful one of complex origin where both tectonics and vulcanism have played an important part. The probable youthful nature of the province is reinforced by the fact that almost all earthquake epicenters within the basin of the Arctic Ocean fall in this province. On the basis of this seismicity, Emery (1949) predicted the presence of a trans-Arctic mountain range which fits the position of this Region of Seamounts. Emery appears to have been correct even though both the seismically calm Lomonosov Ridge and Central Arctic Rise are more impressive trans-Arctic features.

Additional evidence of the volcanic nature of this terrain is provided by the mineral assemblages in the Arctic Ocean Basin: the minerals found west of Lomonosov Ridge indicate derivations from silicic metamorphosed eruptives, whereas those to the east and those in the vicinity of the Region of Seamounts indicate derivation mostly from mafic volcanic eruptives (Saks *et al.*, 1955).

Heezen and Ewing (1961), without knowledge about the presence of the Region of Seamounts as revealed on the SKATE and NAUTILUS echograms, postulated that an extension of the Atlantic Mid-Oceanic Ridge should be present in the Arctic Ocean at the position of the Region of Seamounts. Typical portions of SKATE echograms showing details of the Region of Seamounts are presented in Figure 2 of Plate 4. These and the NAUTILUS profile (Fig. 2) show the striking similarity of this province to portions of the Atlantic Mid-Oceanic Ridge (Heezen *et al.*, 1959). Detailed echo-sounding on the Atlantic Mid-Ocean Ridge has revealed that many features which appear as jagged peaks on cross sections with considerable vertical exaggeration are in reality ridges. If the Region of Seamounts is a continuation of the Mid-Oceanic Ridge, which appears likely to us, then at least some of the peaks shown by the submarine echograms are probably really ridges.

GREENLAND-SPITZBERGEN (NANSEN) RIDGE: The final geomorphic province crossed was Nansen Ridge, which extended for more than 930 km along the track of NAUTILUS. This feature is a broad swell of intermediate depth connecting Greenland and Spitzbergen. It provides the only connection between the Arctic Ocean and the other oceans of the world that is deeper than the depth of the continental shelf. However, it is not the ultimate sill for Arctic Ocean waters because an even shallower barrier is imposed at Denmark Strait between Iceland and Greenland.

On its poleward flank, Nansen Ridge rises gradually from depths of 4000 m to 2600 m. Then it ascends abruptly to a broad crest at depths between 1280 and 1460 m which is about 90 km across, and descends gradually to 2200 m, where an abrupt scarp causes the sea floor to drop from 2200 m to 2570 m. About

33 km farther south, Nansen Ridge is abruptly terminated by a spectacular scarp 1800 m high, which drops the sea floor from 2750 m to 4575 m, and which has a slope of 15°. The base of this scarp is deeper than any sounding recorded by NAUTILUS in the Arctic Ocean. The scarps may mark major crustal fracture zones.

The surface topography of Nansen Ridge is gently undulatory and lacks the small surface irregularities commonly present in volcanic terrains. The over-all smoothness suggests that it is an old structure. The surface relief and dimensions of the ridge suggest a continental nature.

ARCTIC BASIN—OCEAN BASIN OR MEDITERRANEAN BASIN?

Various writers have discussed whether the basin of the Arctic Ocean is a true ocean basin or a mediterranean basin; unfortunately they have had relatively few data to back up their discussions. Eardley (1948) was convinced of its mediterranean nature. From the point of view of physical oceanography, the Arctic Ocean is a mediterranean because it is almost completely landlocked. Nearly all the interchange of water must be effected through Denmark Strait. But from the geological viewpoint, the mediterranean nature of the basin of the Arctic Ocean is a moot question. On one hand, the basin is landlocked by continental barriers, being surrounded on all sides by continental shelves except for Nansen Ridge, and the topography of Nansen Ridge suggests that it is continental material. Further, Lomonosov Ridge ties into the continents on both sides of the basins as though it too were a part of the continental framework. Such extensions of continental structures rarely are found in ocean basins. On the other hand, seismic-refraction studies of all oceans have established the general principle that water deeper than about 3500 m is underlain by oceanic-type crustal structure. Inasmuch as the three basins of the Arctic Ocean all are deeper than 3500 m, it would seem that they are truly "windows" in the continental blocks and are underlain by an oceanic type of crust.

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Quantitative Areal Modal Analysis of Granitic Complexes

Abstract: The difficulty of obtaining an adequate quantitative estimate of the composition and variability of a granitic complex is discussed. Sampling problems are considered in relation to the sampled and target populations. Trend-surface analysis is an efficient method of studying selected areas of the Donegal "older granite." Different levels of sampling are compared. The trend components for several mineralogical variables and the results of regression analysis suggest that one specimen per quarter square mile gives a reasonable approximation of the trend components of the granite; when averaged these data yield comparable trend components more economically than the modes for individual data points. Use of only one specimen per square mile gives spurious results, although with this level of sampling some variates give better results than others (*e.g.*, color index gives better results than total feldspar percentage). The most efficient sampling plan for effective quantitative estimates to be made varies according to the objective of the study. Although averages yield good trend surfaces, they mask important features in the deviation maps. Closely spaced individual data points appear to allow better correlation between systematic deviations and the underlying geologic causes of such local departures (deviations) from the regional trend.

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INTRODUCTION

It is no longer adequate simply to identify and map qualitatively the components of an igneous complex. Examination of recent literature shows a rapidly increasing awareness of the necessity for quantitative data. Discussions of such topics as magmatic differentiation, contamination between magmas, metasomatism, or any aspect of granitization necessarily involve consideration of the quantities of chemical and/or mineralogical components involved (*e.g.*, Nockolds and Mitchell, 1948). Too frequently data are obtained by considering a linear traverse represented by 6–12 specimens, or even by comparison of a set of typical representative samples (*e.g.*, Lapadu-Hargues, 1945).

It is difficult to estimate what significance should be attached to a typical specimen. Presumably, after thorough field study of a complex, a geologist makes a subject-matter decision and selects samples to represent the area. Such specimens may be typical with respect to those properties he can readily estimate but atypical with respect to other properties. Even if one property can be determined readily by laboratory measurement, it is still necessary to know how an investigator can discern whether a sample is typical. Alternatively, if such a specimen is taken from a rock mass which is thought to be homogeneous, what sampling program is necessary to prove the homogeneity? What sampling will establish inhomogeneity adequately? Once inhomogeneity is proven, what sampling program is necessary to define effectively the nature of the variability? Will the same sampling plan be adequate for all variable attributes of a particular complex? Will the same sampling plan be adequate to define the variation pattern of the same property in other rock bodies or complexes?

Typicality and inhomogeneity depend on the scale factor. From within a batholith which extends over many hundreds of square miles, samples from a single square mile might indicate that attributes X_1, X_2, X_3, . . . have a nearly uniform or homogeneous value. However, within this same unit other variables Y_1, Y_2, Y_3, ... may possess an inhomogeneous distribution. By contrast, in sampling the whole batholith different variables Z_1, Z_2, Z_3, ... and, for example, X_2 and Y_3 may be homogeneous throughout the mass, while X_1, X_3, Y_1, Y_2, etc., are inhomogeneous. Thus, over a limited area (*e.g.*, 1 square mile) hornblende or biotite content may be constant or have a simple uniform variation across the area, whereas the quartz content may be inhomogeneous; over the whole batholith the mafic minerals could have a markedly patchy or inhomogeneous distribution, but the quartz content could be essentially constant throughout. On an even smaller scale serial sections cut from a single hand specimen will in many cases yield modes with comparable biotite or hornblende percentages but varied quartz percentages. Thus, in a quantitative study knowledge of the levels of homogeneity of each variable is important in planning the sampling program. A comparable and equally serious problem associated with levels of homogeneity exists in structural geology; in an extensive survey of the problem Weiss (1958) proposed the terms micro-, meso-, and macrostructure with respect to which different structural elements may be homogeneous. A similar consciousness of scale is necessary in ordinary quantitative petrography and petrology.

If a large number of specimens is to be collected from a batholith, it might be assumed that sampling could be efficiently performed by the selection of probability sample locations. Locations on a graticule superimposed on a map could be selected with the aid of a table of random numbers. If specimens are collected according to such a design, one of the standard statistical analytical procedures could be employed. However, because of the vagaries of natural outcrop, the localities chosen will not always be available for sampling; even in well-exposed terrains, it is common to find that large areas are totally unexposed, whereas other areas may show only scattered isolated outcrops.

Geologists have not addressed themselves to the fundamental sampling problems involved. Recently Krumbein (1960) discussed the subject of "target populations" in geological studies and the limitations imposed by the vagaries of natural cover which conceal parts of the whole outcrop; the nature of the "target population" with respect to a granitic batholith is discussed in the following section. The quantitative "target population" frequently seems to be a purely academic question, solely represented in many petrologists' thinking by a "typical specimen," or the average of a few miscellaneous specimens, whose original location is so unimportant that the journal reader is not burdened with the details. Is it realistic

to attempt a petrogenetic reconstruction of a complex without ascertaining the nature and variability of its minerals and recording precisely the relative positions of the specimens studied? Appraisal of any conclusions is made almost impossible when details of the spatial distribution and variability of the specimens studied are not given.

For the purposes of this paper the writer assumes that the modal composition of a hand specimen can be determined within satisfactory limits of accuracy. On the basis of this assumption the broad problems associated with areal sampling are examined. At present a rigorous mathematical treatment of the problems involved in sampling of this type would be beset by many unknown factors which will remain unknown until a considerable volume of suitable quantitative data has been amassed and studied. In consequence, many decisions related to the suitability of both the sampling programs and the results obtained from analysis of data must be based on geological subject-matter experience, because no formal test can yet be erected. An adequate sampling plan will provide an unbiased estimate of the population under study, provided that the requirements of such a plan can be met in practice.

This paper embodies the results of some preliminary experiments. Rather than providing answers to the questions posed above, it exposes certain aspects of the problem. The experimental results indicate the type of sample data which seem to be necessary to derive valid quantitative conclusions. The study shows that the method and density of sampling necessary depends on so many factors that results obtained for one granitic massif may not be directly applicable to another, although the writer anticipates that the general method, which involves trend-surface analysis, will be found suitable for many two- or three-dimensional variation problems.

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TARGET AND SAMPLED POPULATIONS

The term "target population" has a technical meaning in statistical usage (*e.g.*, Cochran *et al.*, 1954) and has recently been introduced into the geologic literature (Krumbein, 1960). The target population is the population of specimens of interest, about which the geologist wishes to make inferences or draw conclusions. In a batholith this population is the object of study and comprises all the potential specimens constituting the uppermost layer of the batholith parallel to the eroded surface. Hence, the target population is a vast, but finite, number of hand specimens. In a complete three-dimensional study, all the equal-sized hand specimens potentially constituting the entire batholith would be the target population. Some features of both the target and sampled populations are illustrated in Figure 1. In Figure 1A some areas are completely obscured by bog, gravels, and a lake, whereas areas of almost continuous outcrop are specifically outlined. As a basis of generalizations about the whole batholith (at the present level of exposure), the target population of interest comprises all the N hand specimens composing the surface of the granite enscribed by the dashed line. However, the bog, gravels, and lake cause part of the target population to be inaccessible for all practical purposes. Specimens are therefore drawn from actual exposures, and all the n hand specimens composing the entire surface of these natural and artificial exposures constitute the "sampled population."[2] In an example like that represented in Figure 1A, collection of the

[1] A program prepared by W. C. Krumbein for the Standard IBM 650 which computes trend-surfaces up to the cubic for irregularly spaced data points is available in the library of the Data Processing Division of International Business Machines Corporation, New York.

[2] It is understood that Griffiths (In press) intends to subdivide the target population of Cochran *et al.* (1954) into the hypothetical, existent, and available populations, of which the latter equates with the present sampled population.

n specimens of the sampled population would be an impossible task in practice, but it is only by study of the sampled population that geological inferences can be made concerning the target population, which is the real subject of interest.

If a rectangular grid is laid out across a batholith, specimens can be obtained only from a limited number of grid intersections owing to the limitations of natural outcrops (Fig. 1A). In consequence, there are severe restrictions on direct probability sampling from the target

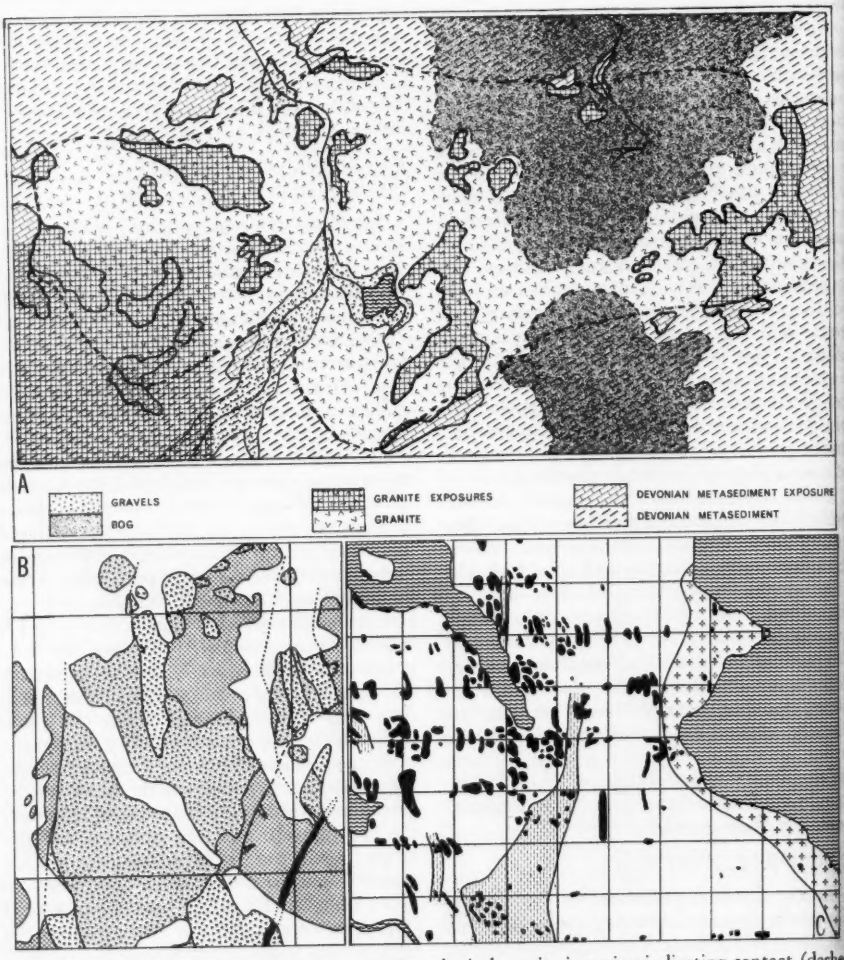

Figure 1. Target and sampled populations. A. Hypothetical granite intrusion indicating contact (dashed line) between granite and Devonian metasedimentary rocks. A 500-foot grid is symbolically represented at the bottom corner of the map; this grid should continue across the whole map, but it is indicated only on exposed areas of the granite. B. Detailed map of part of the bronzitite-harzburgite-dunite complex in the Blakely Creek area, Sweetgrass County, Montana (after Howland *et al.*, 1949, Pl. 38). Blank area is unexposed; solid black is a diabase dike; other patterns indicate where various ultramafic igneous rocks outcrop. Grid lines are 500 feet apart. C. Detailed map showing observed rock exposures in part of T. 63 N., R. 11 W., Minnesota (after Clements, 1903, sheet 22). Observed rock exposures—solid black; Ely greenstone—blank; Soudan formation—dashed lines; Giants Range granite—crosses; water—ripple lines. Grid lines are a quarter of a mile apart.

population, but it would be possible to take a probability sample from the hand specimens composing each area of continuous outcrop (the sampled population).

In practice the sampled population comprises the potential hand specimens forming the surface of all the actual outcrops which the geologist succeeds in visiting and sampling; he may not visit all outcrops, and the specimens that constitute the surface of the unvisited exposures are not part of the sampled population. Figure 1C shows the exposures actually visited and mapped in a small area of Minnesota (Clements, 1903); in this case the sampled population was governed by the actual traverses made in the field as well as the vagaries of superficial cover, such as lakes and marshes. From these samples inferences were drawn about the target population—*e.g.*, the position of contacts between the rock types.

In a sampled population related to the near-surface part of a series of sand dunes, a grid could be laid out and sampling locations chosen from random-number tables; specimens could probably be collected from all the designated localities. The result would be probability sampling, because the chance that every sampling unit has of entering the gathered sample is prescribed. In collecting granite specimens a grid can also be laid out with respect to the sampled population (Fig. 1A), but in practice the geologist does not have complete freedom to collect all the prescribed specimens. Usually upper semiprobability sampling is achieved, since the grid squares (or intersections) can be located at random, although the particular hand specimen of granite will not be chosen at random. Factors such as ease of collection and lack of weathering will control the choice of actual specimens taken from a prescribed locality (*cf.* Cochran *et al.*, 1954, p. 294, 322.)

Figure 1B illustrates an actual outcrop pattern of some ultramafic rocks in the Blakely Creek area of Montana (after Howland *et al.* 1949), and the grid shown was that used in the original survey. On the scale of Figure 1A the area of Figure 1B would be continuous exposure, but collecting specimens at the grid intersections would be difficult; at three intersections exposures do not occur, and the fourth intersection coincides with a later dike. Hence, for purely practical reasons hand specimens often have to be collected from subjectively selected sites as near as possible to prescribed grid locations.

In favorable circumstances it is possible to make rigorous statistical inferences about the sampled population; then the geologist may make subject-matter inferences about the target population based on the experience gained from the sampled population.[3] For example, the trend components discussed below relate to mineralogical variables, and the partial-trend surface for each variable approximates the complete trend for that variable within the sampled population. By subject-matter geological reasoning an attempt is made to draw inferences about the behavior of each variable in the target population. "The step from sampled population to target population is based on subject-matter knowledge and skill, general information, and intuition—but not on statistical methodology" (Cochran *et al.*, 1954, p. 19).

The target population must be carefully defined for each investigation, and in each case the precise nature of the target population is likely to be considerably different. For example, the target population might comprise all the hand specimens potentially available from the layer of granite immediately beneath the eroded surface of a 25-square-mile portion of a batholith. By contrast, the target population might comprise all the specimens 2 by 2 by 2 inches contained in the 2-inch-thick sheet at 750 feet above sea level (at the ground surface, underground, and already eroded away) of the Newer Granites of Scotland; these granites constitute a number of granitic outcrops unconnected at the present level of exposure.

METHODS OF STUDY

Attempts to assess the quantitative areal variability of granitic rocks by the use of various methods have been made by Gross (1950; 1952), Saha (1954; 1958; 1959), Hamilton (1956a; 1956b), Taubeneck (1957), Dawson (1958), Ross (1958), Mehnert (1960), and Mercy (1960a; 1960b). In these papers varying success was obtained in estimating the areal and/or linear modal and chemical variability. Following earlier variation studies (Whitten, 1957), the writer (1959a; 1959b; 1960) showed that trend-surface analysis provides a valuable means of expressing quantitative mineralogical variation. The optimum sampling design for a particular complex depends *inter alia*, on (1) the specific aspect of variability under investi-

[3] M. A. Rosenfeld (1953, Ph.D. thesis, Pa. State Univ.) provided an example, based on the size of quartz grains in the Oriskany "Sandstone Complex."

gation—*i.e.*, whether trend-surface components or deviation patterns are involved—and (2) the variable involved.

The method by which partial-trend surfaces are fitted to a set of modal data has been outlined (Whitten, 1959a). In essence linear and quadratic (or higher-degree) surfaces are fitted to the observed modal data by least-squares techniques. Grant (1957) gave the mathematical background to trend-surface analysis. Krumbein (1959a) extended the analysis to include irregularly spaced control points and referred to the "trend component" as a particular selected part (*e.g.*, linear partial-trend surface) of the complete trend defined by Grant.

Deviation maps contain the residual terms of Grant (1957) and some trend terms of higher degree than the trend component used; the deviations are the differences between observed values at any map point (for example, modal composition) and the values at that map location on the particular computed partial-trend surfaces. Examples of both partial-trend surface and deviation maps have been illustrated by Whitten (1959a; 1960). Such maps, which result from polynomial analysis of each mapped variable, give useful approximations to the actual composition and areal variation of a rock body.

The technique is most effective and accurate when the control points (specimen localities) are equally spaced on a rectangular grid extending over the entire area of study. Such a specimen distribution enables orthogonal polynomial analysis to be used. In contrast with orthogonal analysis, the available methods of nonorthogonal analysis allow only a limited number of trend components to be calculated. Commonly in sampling an igneous complex it is impossible to lay out a grid and to collect specimens at predetermined grid points (*see* above); in consequence, modal data usually relate to irregularly spaced localities, and nonorthogonal polynomial analysis is employed. (*See* Krumbein, 1959a.)

In the present case the mapped variables have been derived from modal analyses, although many other parameters, such as bulk chemical composition and specific gravity, could be used. In the following discussion maps showing the variability of the following seven variables are given: (1) X_1—quartz percentage, (2) X_2—color index (as used by Shand (*e.g.*, 1947) in relation to the ratio of heavy to light minerals in an igneous rock), (3) X_3—ratio of potassic feldspar to plagioclase, (4) X_4—feldspar percentage, (5) X_5—epidote percentage, (6) X_6—hornblende percentage, (7) X_7—sphene percentage.

In effect color index comprises all the minerals except quartz and feldspar, so that variables (1), (2), and (4) always total 100 per cent, and

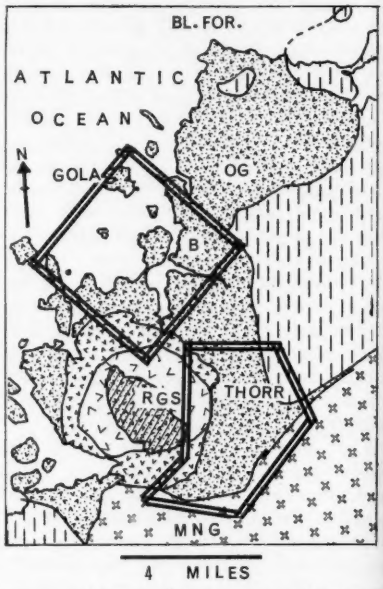

Figure 2. Index map of the portions of the Donegal granites (northwest Ireland) studied. The Bunbeg area and Thorr district are enscribed. OG—"older granite" of Donegal; RGS—Rosses granitic ring-complex; MNG—Main "younger granite;" BL. FOR.—Bloody Foreland; B—Bunbeg; vertical rules—metasedimentary rocks

(5), (6), and (7) constitute part of (2). Secondary alteration products of the feldspar have been ignored.

Additional variables based on modal data could be profitably studied. Certain difficulties are associated with statistical treatment of percentage data (*e.g.*, Chayes, 1948; 1960; Vistelius and Sarmanov, 1961), and possibly various ratios might give preferable results. The use of modal data not converted to percentage scale has not been possible in the present study. All the modes were determined with a Swift automatic electrical point counter, with points 1/3 mm apart in traverses spaced at 1 mm.

SIMPLE RELATIONSHIPS IN REGIONAL VARIATION

To illustrate some of the problems involved, the quantitative mineralogy of 25 square miles of the "older granite" of Donegal is examined; the area studied includes the small hamlet of Bunbeg, which lies on the Atlantic seaboard of northwest County Donegal, Eire (Fig. 2). Inferences concerning the Bunbeg area are compared with results derived from the Thorr district of the same granitoid mass. The granite is a medium-grained Caledonian mass enveloped by Dalradian (Precambrian) metasedimentary rocks.

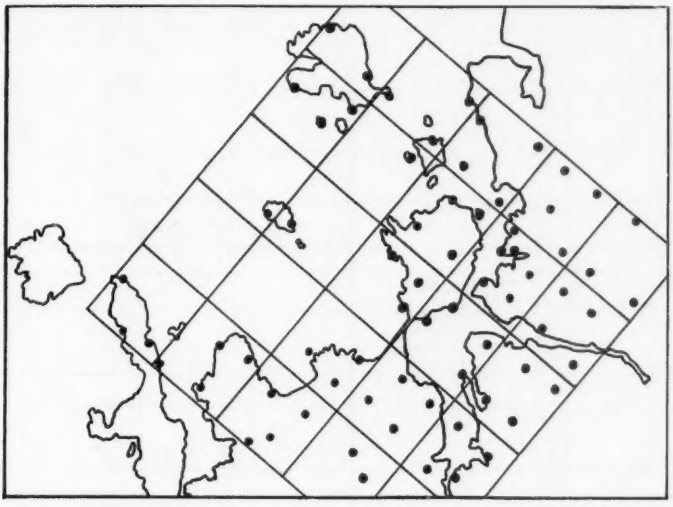

Figure 3. Locations of all specimens utilized in Bunbeg area. Each grid is 1 square mile.

Within the Bunbeg area the four variables, quartz percentage, color index, ratio of potassic feldspar to plagioclase, and total feldspar percentage have been studied. For this area 71 modes are available (Fig. 3), and the specimens are related to a square-mile grid subdivided into quarter square miles. A mode is available for every quarter square mile in which outcrops occur. The scale of the maps makes it impossible to indicate all the innumerable rocks and islets in the ocean; several of the specimens come from such exposures, but 3 square miles are devoid of outcrops. Originally a few specimens were collected randomly, according to no statistical design, as typical specimens; subsequently, after laying out the grid, specimens were collected (for unrepresented quarter square miles) from as near the center of each square as exposure and collecting conditions permitted. The actual modal data and their locations are given in Table 1.

Two levels of sampling are compared for each of the four variables. In experiment 040001 linear partial-trend surfaces have been computed on the basis of one specimen (Fig. 4A) per square mile, that in the western quarter square (or, if that is vacant, the first specimen in west-north-east-south sequence). The four partial-trend surfaces are represented in Figure 4C by contours drawn so that the distance between each is proportional to the spacing of 1 per cent contours on the surface (except for feldspar ratio, where the contour interval is proportional to ratio differences of 0.1). In a second experiment (040002), illustrated in Figures 4B and D, similar surfaces are based on the average of all modal data within each square mile of the grid.

Comparison of Figures 4C and D reveals marked differences. The four partial-trend surfaces of Figure 4D (experiment 040002) are essentially parallel; although their dip varies a little, the strike, which is more significant, is essentially parallel for each variable. In experiment 040001 (Fig. 4C) the range of dip is the

TABLE 1. MODAL DATA FOR BUNBEG AREA, COUNTY DONEGAL, IRELAND

u		v	Quartz percentage	Color index	K-feldspar / Plagioclase	Total feldspar percentage	Points*
2.45	a	7.10	31.49	1.93	1.21	66.58	2807
2.15		7.60	24.25	4.21	1.05	71.54	2091
2.80	c	7.20	28.64	1.81	1.17	69.55	3373
2.85**	ac	9.25	39.45	1.48	0.79	59.07	2441
2.22	a	10.51	29.28	0.95	1.25	69.77	3275
2.90		10.80	23.54	1.13	1.08	75.33	1682
2.70	c	10.40	54.06	1.16	1.12	44.78	1639
2.25	a	11.45	33.71	0.68	1.03	65.61	1925
2.75	c	11.20	46.67	1.04	1.24	52.29	1920
3.10	a	7.10	38.40	1.97	0.74	59.63	2026
3.47		7.70	32.68	8.32	0.44	59.00	1983
3.80		7.77	23.90	4.73	0.79	71.39	1554
3.60	c	7.20	17.25	5.61	1.25	77.20	1623
3.21**	ac	9.30	33.15	1.03	0.97	65.82	1653
3.77	ac	10.77	34.91	2.71	1.73	62.39	1845
3.10	a	11.22	25.21	0.95	1.07	73.83	4720
3.78		11.90	26.68	3.81	0.70	69.51	3362
3.80	c	11.20	39.03	2.62	1.68	58.35	1868
4.45	a	7.10	17.06	8.98	0.30	73.96	1905
4.30		7.70	28.95	5.73	0.76	65.32	1606
4.75		7.75	22.85	7.46	0.76	69.69	1891
4.62	c	7.25	25.89	8.61	0.92	65.51	1580
4.35	a	8.30	33.94	8.48	0.85	57.58	1438
4.80		8.65	15.33	8.71	1.03	76.33	1677
4.77	c	8.24	16.86	10.35	0.46	72.78	1951
4.23	a	9.80	21.35	7.65	0.55	71.01	2197
4.70		9.68	27.98	1.98	1.53	70.04	1619
4.80	c	9.40	12.52	3.70	1.19	83.77	2052
4.25	a	10.25	24.58	2.06	1.25	73.36	1554
4.44		10.65	14.87	6.33	1.78	78.80	1486
4.75		10.72	28.96	4.86	0.71	66.18	1789
4.85	c	10.25	44.42	8.54	0.44	47.04	1522
4.30	a	11.20	30.60	1.59	1.83	67.81	2575
4.20		11.70	49.84	1.02	0.79	49.14	1864
4.80		11.90	9.02	3.43	0.94	87.56	2041
4.90	c	11.10	14.80	12.41	0.36	72.79	1547
5.48	a	7.76	12.89	9.48	0.49	77.64	3671
5.80	c	7.70	28.16	7.91	0.84	63.93	1630
5.20	a	8.35	18.18	15.07	0.80	66.75	1612
5.30		8.79	15.87	14.05	0.20	70.08	2256
5.84		8.75	21.73	7.08	0.89	71.19	1482
5.62	c	8.25	19.13	9.07	0.56	71.80	1532
5.20	a	9.50	35.13	7.60	0.26	57.27	1987
5.22		9.75	34.73	6.42	0.61	58.85	2197
5.85		9.60	16.22	16.36	0.64	67.43	1432
5.85	c	9.25	9.17	9.11	0.82	81.73	2052
5.30	a	10.25	15.27	7.95	0.75	76.78	1572
5.25		10.94	25.86	8.71	0.40	65.43	1802
5.27		10.65	21.32	8.14	0.27	70.54	1843
5.41		10.78	37.77	13.25	0.35	48.98	2007
5.77		10.70	21.03	5.74	1.06	73.24	3139
5.75	c	10.30	23.96	3.31	0.66	72.74	2087
5.40	a	11.40	20.25	5.15	0.59	74.61	1457
5.21		11.80	22.42	8.10	1.19	69.48	1494
5.75		11.90	27.66	5.13	0.61	67.21	2202
5.87	c	11.12	19.11	16.76	0.50	64.02	1826
6.25	a	8.20	23.95	10.35	0.10	65.84	1411
6.20		8.75	26.18	13.38	0.32	60.44	2399
6.73		8.75	25.42	10.70	0.68	63.88	2655
6.60	c	8.35	11.28	9.44	0.55	79.29	1685
6.26	a	9.25	20.84	11.86	0.73	67.31	1771
6.41		9.77	25.70	10.25	0.43	64.06	4265
6.78		9.88	17.04	14.18	0.52	68.78	3626
6.65	c	9.24	6.95	14.44	0.33	78.61	2071
6.22	a	10.25	21.55	7.84	0.96	70.62	1889
6.22		10.76	25.48	8.93	0.37	65.59	1523
6.58		10.77	24.97	10.84	0.43	64.19	1790
6.72	c	10.20	23.62	12.56	0.31	63.82	1744
6.27	a	11.12	26.69	3.04	0.71	70.27	2136
6.28		11.95	18.69	9.20	0.49	72.11	2001
6.79	c	11.22	23.27	11.53	0.77	65.21	1414

* Point count traverse length

** These data averaged for experiment 040002

a = specimens used for experiment 040001

c = specimens used for experiment 040003

u and v are expressed in units of 1 mile, with the origin to the southwest of the Bunbeg area. The westernmost corner of the Bunbeg area is at u = 2.000, v = 7.000, and u increases southeastward and v northeastward; horizontal lines separate data within each square-mile grid square.

same, but the strikes of the surfaces for the different variables are much more diverse. Subject-matter geological experience, derived from a thorough field investigation of the region, reveals that there are definite mineralogical gradients within the granite which have a common locus northwest of the present area. This suggests that there is a continuous gradual transition on a regional scale, and geological experience suggests that each mineralogical phase probably participates in such a gradation. Hence, on essentially empirical grounds Figure 4D appears to yield the more realistic portrayal of the complete trend of the four variables within the target population (for Bunbeg area). This conclusion is strengthened by the fact that arbitrary selection of a different set of single specimens per square mile yields partial-trend surface strikes just as diverse as those shown in Figure 4C.

This last assertion, namely that different selections of single specimens per square mile yield diverse results, is emphasized by Figure 5. In order to illustrate the point, in this figure an arbitrary selection of linear and linear-plus quadratic partial-trend surfaces for three variables was made. All six surfaces are based on 22 data points—1 per square mile; to provide a contrast with the surfaces developed in experiment 040001 (Fig. 5, upper row) the specimen in each of the southernmost quarter square miles (or if vacant the first specimen in a southeast-north-west sequence) is utilized (experiment 040003) in the lower row of surfaces (Fig. 5). In Figure 5 the proportion of the total sum of squares associated with each surface is

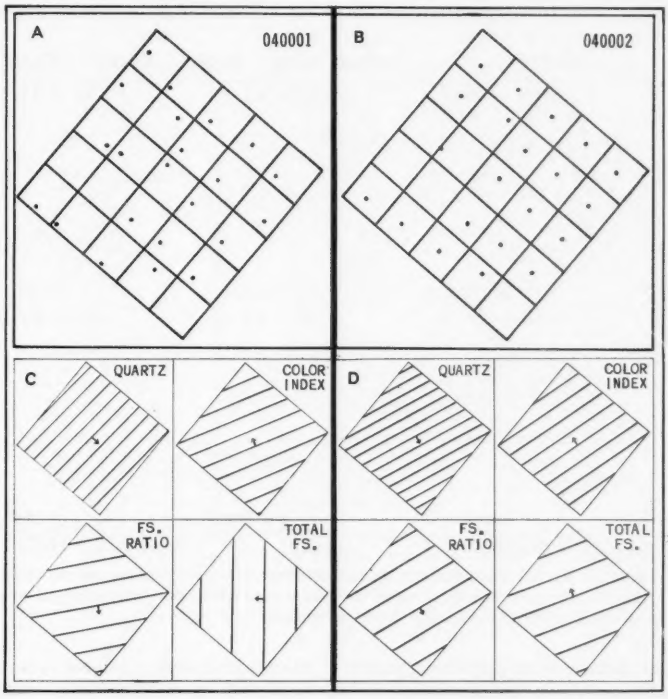

Figure 4. Comparative linear partial-trend surfaces for quartz percentage, color index, feldspar ratio, and total feldspar percentage in Bunbeg area. Upper diagrams (A and B) show locations of data points on which surfaces below are based. A. Location of the one mode in each square mile on which surfaces in C are based. B. Location of the average modal compositions on which the surfaces in D are based. Note that the surfaces show dissimilar strikes in C and similar strikes in D.

indicated; it has been argued previously (Whitten, 1959a, p. 838) that the sum of squares is a useful criterion in assessing the importance of a surface in relation to the observed data and in comparing the partial-trend surfaces.

Figure 5 shows that partial-trend surfaces for color index account for a much greater proportion of their total sums of squares than do either the quartz-percentage surfaces or the feldspar-ratio surfaces. Thus, the linear partial-trend surfaces for color index account for 65–70 per cent of their total sums of squares (69.2 per cent for 040001 and 65.4 per cent for 040003), whereas the higher-degree linear-plus quadratic surfaces for quartz account for only 46–65 per cent, and for microcline-plagioclase ratio 45–51 per cent. Although it appears that the color-index surfaces are highly significant in relation to the data used for each analysis, the difference in strike of the two surfaces is marked. In both cases the linear surfaces are significant with respect to the particular set of 22 modes employed but show discordance from the regional trend implied by Figure 4D and thus presumably from the complete trend of the target population. Thus, the two color-index surfaces of Figure 5 are discordant because of local variability (departure from the regional complete trend) of the particular samples employed.

When the linear-plus quadratic partial-trend surfaces for quartz percentage and microcline-plagioclase ratio are considered (Fig. 5) the discrepancies resulting from the two sampling schemes (040001 and 040003) are more blatant. Although higher-degree partial-trend surfaces are figured, the proportion of the total sum of squares accounted for in each is surprisingly small; this implies a considerable deviation of the observed data from the computed surface in each case, and that local variability is large compared to the linear-plus quadratic trend components. The two partial-trend surfaces for quartz percentage show completely different trends; each surface represents real trend components with respect to the set of data employed but is spurious as an estimate of the regional trend of the target population.

The feldspar-ratio diagrams (Fig. 5) imply that a similar situation exists for that variable. From such results it is difficult to draw any conclusions about the trend components for each variable of the target population. However, Figure 4D suggests that trend components based on the average data reflect the trends of the target population. In terms of homogeneity discussed above, both quartz percentage and feldspar ratio are inhomogeneous at the local level because of local variability unconnected with the trend of the target popu-

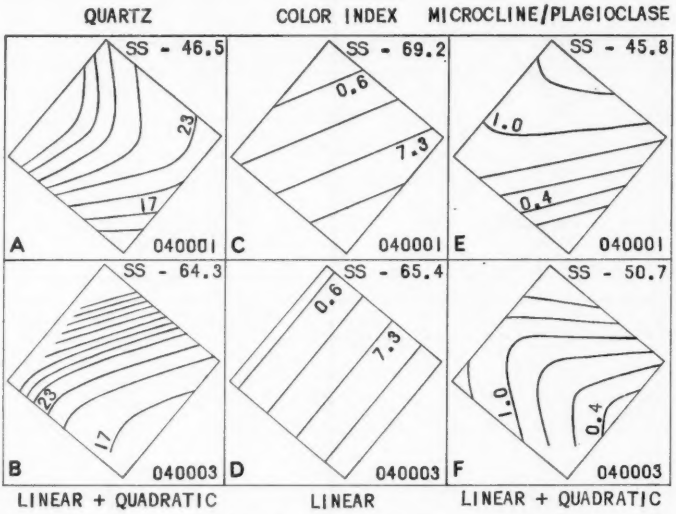

Figure 5. Partial-trend surfaces for the Bunbeg area for comparison of two suites of one specimen per square mile (040001 and 040003). SS—percentage of sum of squares accounted for by the trend surface

lation, whereas at the regional level (with respect to the Bunbeg area) the variability is apparently (empirically) conformable with that of the other variables. The above results indicate that conclusions based on one mode per square mile are erratic. With ranges of the general order shown in Tables 1 and 2 it would be astonishing if this were not so. These results may help to establish how imperative it is to use an adequate sampling scheme before assuming that any particular quantitative results are meaningful.

The consistent variability pattern shown by each of the variables on the regional scale is assayed by means of averaged data in Figure 4D. The use of averages is not always desirable, and in geological studies, a false impression of uniformity can be obtained by "averaging out" divergent data. It is instructive to compare and contrast partial-trend surfaces based on averages (Fig. 4D) and all the available individual modes for the same area. Comparative pairs of such surfaces for color index, feldspar ratio, and sphene percentage are illustrated in Figure 6; similar results for the same area are obtained for quartz, total feldspar, hornblende, and epidote percentages (Fig. 7). The region illustrated in Figure 6 is Thorr district, which lies a few miles southeast of Bunbeg (Fig. 2), and the modal data are given in Table 2.

The three lower diagrams of Figure 6 are based on the individual modes in each quarter square mile (65 data points—experiment 040011); the upper diagrams relate to the averages of the same data within each square mile of the grid (21 data points—experiment 040012). The grid is a continuation of that used for Figures 4 and 5. Figure 6 indicates that the linear-plus quadratic partial-trend surface appropriate to both sampling schemes has essentially the same geometric form for each variable. It is difficult to erect criteria of geometric likeness for comparing such trend surfaces, but the over-all shape of the surfaces for each variable and the skewness of the bisectrices of the contours shown in Figure 6 are remarkably similar. Geometrically all the surfaces illustrated in Figure 6 are strikingly similar, although very dissimilar variables are compared and the absolute numerical values involved are very different.[4]

The proportion of the total sum of squares accounted for by each surface is indicated in Figure 6. Averages yield surfaces which are not

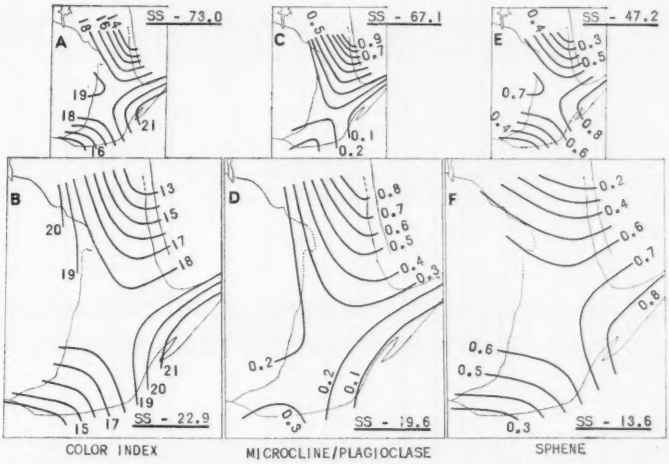

Figure 6. Comparative linear-plus quadratic partial-trend surfaces for the "older granite" of Thorr district. The three upper surfaces are based on averages of data for four specimen localities per square mile (experiment 040012); the three lower surfaces are based on one mode per quarter square mile (experiment 040011). SS—percentage of sum of squares accounted for by the trend surface

[4] Although such comparison may seem subjective, the problem is presently being studied quantitatively by W. C. Krumbein and the author, and there is good reason to believe that such surfaces are susceptible to objective comparison.

TABLE 2. MODAL DATA FOR THORR DISTRICT, COUNTY DONEGAL, IRELAND

U	V	Quartz percentage	Color index	K-feldspar / Plagioclase	Total feldspar percentage	Hornblende percentage	Sphene percentage	Epidote percentage	Points*
7.84	8.35	23.34	21.69	0.27	54.98	1.87	0.30	0.43	2301
8.20	8.28	17.72	18.91	0.64	63.37	0.05	0.64	2.72	2020
8.30	8.75	15.29	17.36	0.11	67.36	2.71	0.32	0.77	1549
8.77	8.80	18.09	11.73	0.31	70.18	0.85	0.64	0.35	1415
8.70	8.38	20.83	25.60	0.14	53.57	3.36	0.97	0.33	1848
8.75	9.40	21.75	13.97	0.40	64.28	0.74	0.25	0.80	1632
9.90	6.95	14.89	18.47	0.33	66.63	1.06	0.86	0.51	1981
9.45	7.96	11.89	20.73	0.38	67.37	4.42	0.44	0.60	1833
9.61	7.82	16.50	17.80	0.53	65.69	1.16	0.33	0.42	2151
9.55	7.35	15.88	15.20	0.14	68.93	0.50	-	-	2198
9.21	8.30	12.48	14.89	0.21	72.63	2.12	0.47	0.88	1699
9.38	8.80	14.73	24.92	0.50	60.35	1.00	0.29	0.53	2091
9.98	8.70	10.14	15.90	0.60	73.96	0.52	0.80	1.04	2120
9.60	8.20	11.97	15.72	0.40	72.31	1.33	0.95	1.80	3167
9.42	9.10	28.68	12.10	0.16	59.22	0.55	0.07	0.89	1454
9.15	9.74	9.87	11.49	0.69	78.65	0.78	0.42	1.34	1419
9.71	9.27	12.97	12.33	2.87	74.90	1.06	0.07	0.21	1419
10.35	3.95	28.66	15.35	0.00	55.99	-	0.04	0.27	2561
10.57	3.90	18.81	15.30	0.57	65.89	-	0.52	1.18	2111
10.40	4.25	35.70	12.33	0.03	51.97	-	0.04	-	2280
10.80	4.80	7.78	18.85	0.12	73.37	-	0.60	1.80	1671
10.61	4.12	34.59	14.93	0.05	50.48	-	0.13	0.04	2304
10.40	5.88	20.63	19.07	0.44	60.30	-	0.84	0.97	2370
10.88	5.80	17.99	17.85	0.25	64.16	0.05	0.36	1.85	2218
10.83	5.23	14.57	24.17	0.32	60.26	-	1.15	2.17	1477
10.24	6.27	21.52	14.55	0.35	63.92	0.04	0.94	0.51	2769
10.65	6.57	27.06	14.92	0.18	57.91	-	-	-	1870
10.58	6.87	21.07	22.84	0.32	56.09	3.10	0.41	0.86	1970
10.94	6.81	11.69	18.01	0.56	70.29	1.73	1.28	0.66	2259
10.90	6.50	36.68	15.61	0.17	47.71	-	-	-	2377
10.27	7.30	18.99	18.74	0.35	62.26	-	1.31	2.97	1990
10.18	7.80	34.20	17.07	0.28	48.74	0.12	0.06	0.06	1623
10.62	7.86	14.53	21.66	0.23	63.82	1.55	0.43	0.73	2327
10.85	7.21	17.60	17.13	0.43	65.28	1.39	0.69	1.39	2598
10.20	8.03	13.46	15.61	0.23	70.93	1.14	0.47	0.74	1493
10.22	8.55	10.01	13.23	0.11	76.76	2.44	0.78	0.78	1493

10.85	7.21	17.60	17.13	0.43	65.28	1.39			
10.20	8.03	13.46	15.61	0.23	70.93	1.14	0.47	0.74	1493
10.22	8.55	10.01	13.23	0.11	76.76	2.44	0.79	0.40	1519
10.70	8.53	26.01	21.47	0.22	52.53	6.57	0.40	0.25	1980
10.85	8.30	29.87	17.99	0.10	52.14	-	0.05	-	2079
10.08	9.05	9.49	15.93	0.74	74.58	3.52	0.35	0.23	1707
11.10	4.27	14.96	11.42	0.62	73.62	0.19	1.30	0.53	2085
11.25	4.80	13.60	17.59	0.42	68.81	-	0.09	0.14	2132
11.70	4.80	29.39	18.16	0.19	52.46	-	-	-	1487
11.15	5.30	20.27	16.57	0.31	63.16	3.80	0.77	1.28	1949
11.36	5.63	16.82	19.88	0.25	63.31	-	1.18	0.47	2123
11.80	5.60	21.03	16.39	0.00	62.58	-	-	-	1812
11.74	5.36	23.99	12.05	0.12	63.96	1.01	0.84	0.36	1670
11.25	6.08	24.25	15.27	0.16	60.48	-	-	-	1670
11.30	6.67	14.59	15.67	0.23	69.74	-	0.54	3.42	2049
11.63	6.66	32.19	24.31	0.09	43.50	0.46	1.25	1.51	1522
11.70	6.36	13.88	14.03	0.64	72.09	0.80	1.15	0.80	1996
11.22	7.24	6.53	19.65	0.59	73.82	3.80	0.79	3.89	2158
11.05	7.80	9.58	23.03	0.07	67.39	4.33	1.25	1.43	1754
11.90	7.91	6.84	24.32	0.04	68.84	2.86	1.55	3.37	2134
11.65	7.30	18.03	18.89	0.14	63.09	2.46	0.86	1.02	2441
11.41	8.13	23.29	17.77	0.02	58.94	1.49	0.80	1.86	1885
11.90	8.70	29.04	15.28	0.37	55.68	2.82	0.74	0.09	2304
11.87	8.03	6.80	26.31	0.01	66.89	5.06	0.42	1.79	1897
12.18	5.40	20.21	21.70	0.04	58.10	-	0.13	0.13	1544
12.28	5.63	28.01	15.98	0.06	56.01	-	0.05	0.09	2146
12.30	6.40	18.29	20.78	0.01	60.93	-	0.83	0.65	1684
12.05	6.85	12.82	24.19	0.22	62.99	1.50	1.17	2.40	2129
12.24	7.25	15.42	10.62	0.43	73.96	2.51	0.17	1.14	1751
12.15	7.85	8.68	29.88	0.02	61.44	0.27	1.15	3.81	2259
12.18	8.30	8.19	22.78	0.01	69.03	1.30	1.00	1.65	2002
12.32	8.65	17.77	16.66	0.27	65.57	1.41	0.72	0.07	3050

* Point count traverse length

U and V are expressed in units of 1 mile, with the same origin as used for Table 1; horizontal lines separate data within each square-mile grid square.

necessarily associated with higher confidence levels, but which automatically account for a much higher proportion of the total sum of squares. However, the form of the upper row of surfaces commends them as indicators of regional trend.

DATA DENSITY NECESSARY FOR QUANTITATIVE MODAL ANALYSIS

In the studies outlined in this paper the basic objective has been to provide an adequate quantitative statement concerning the mineralogical composition and variation of the "older granite" of Donegal to serve as a foundation for subsequent petrogenetic discussion. In this regard the statistical devices employed are essentially means to an end (rather than ends in themselves), and an adjunct to thorough field investigation. The latter must remain the fundamental basis for geologic studies, provided that it is coupled with rigorous sampling procedures when quantitative results are desired. The writer believes that trend-surface analysis provides a realistic

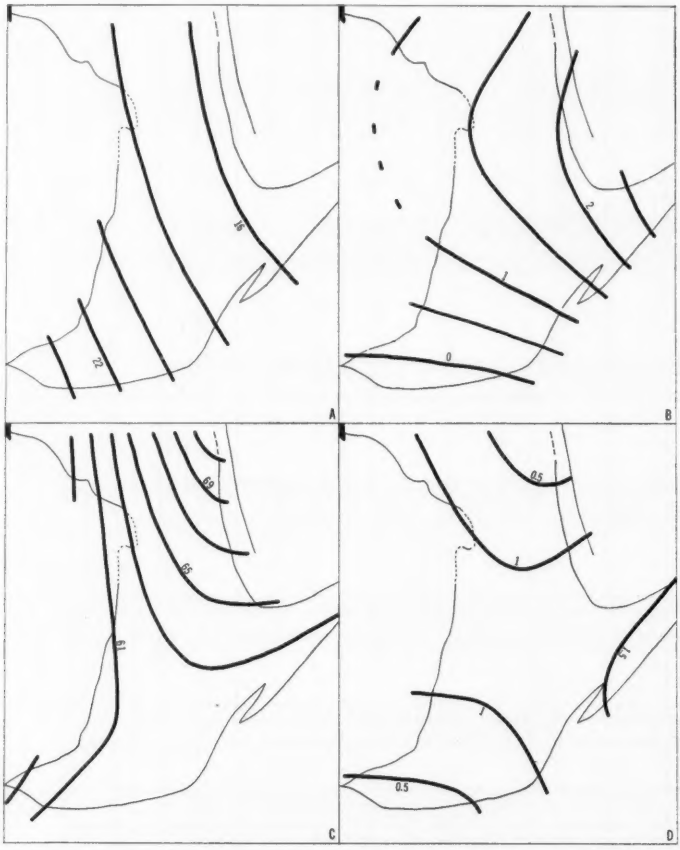

Figure 7. Linear-plus quadratic partial-trend surfaces based on individual data for the "older granite" of Donegal in Thorr district (experiment 040011). A. Quartz-percentage surface, which accounts for 11.8 per cent of the sum of squares. B. Hornblende-percentage surface, which accounts for 23.8 per cent of the sum of squares. C. Total feldspar-percentage surface, which accounts for 10.6 per cent of the sum of squares. D. Epidote-percentage surface, which accounts for 13.6 per cent of the sum of squares

method of portraying the quantitative characteristics of a rock mass.

Several problems have emerged from the study, and their solution is an urgent subject for further research. It is also important to differentiate between (1) geologic subject-matter decision based on field experience and mathematical (and other) analysis of the data collected, and (2) statistical sampling and analytical techniques, and the statistical confidence associated with data irrespective of subjective geological reasoning and experience.

The problems of sampling are of paramount importance and the necessity of adequate sampling schemes cannot be overemphasized. Assuming that the sampling program is satisfactory, and that trend-surface analysis is the best method available for illustrating quantitative composition of the sampled population, it is desirable to investigate the confidence level to be associated with each surface computed for the available data.

As demonstrated above one specimen per square mile is an inadequate data density (Fig. 5) for the Bunbeg area. Although four specimens per square mile appear to provide preferable results, it is unfortunate that, because these analyses are based on nonorthogonal data points, it is difficult to arrive at any estimate of correctness or reproducibility of a particular surface. The best[5] mathematical estimate of significance of a particular trend surface presently available is based on the percentage reduction in the total sum of squares. This estimate is based on:

$$100 \times \frac{\Sigma X^2_{\text{comp}} - \dfrac{(\Sigma X_{\text{comp}})^2}{N}}{\Sigma X^2_{\text{obs}} - \dfrac{(\Sigma X_{\text{obs}})^2}{N}} \% \quad (1)$$

where X_{obs} is the observed value of X, X_{comp} is the computed value of X for the partial-trend surface, and N is the number of data points employed.

These values are used throughout this paper, but in some cases it may be found preferable to base a measure on the $\Sigma\ (X_{\text{comp}} - X_{\text{obs}})^2$, or to adjust the value according to the size of N. If partial-trend surfaces with similar levels of confidence have been computed for areas A (where $N = 25$) and B (where $N = 125$), the percentage reduction in the sum of squares (expression 1) will be much greater for area A than for area B, because $N_B \gg N_A$. An analogous situation arises where average data are employed in place of the individual data (Fig. 6).

Evaluation of the reduction in the total sum of squares is further complicated when a trend component is well approximated by a computed surface, but it represents only a small part of the total value of X_{obs}. In this case the sum of squares will be small, although the trend component computed may be excellent. Such a situation can be visualized readily—*e.g.*, a hypothetical highly variable series of metasedimentary rocks is subjected to metasomatism which causes introduction of 20 per cent quartz in the west and only 5 per cent quartz in the east. Although the original gneisses contained an average of 35 per cent quartz, it was very inhomogeneously distributed. Following metasomatism the resultant rock possesses a definite linear-trend component, although it is considerably masked by the background of inherited variability (noise). In natural problems there is every gradation, from such a case to one in which there is a very strong geological trend which is masked in the modal data only by experimental and operator errors inherent in collection of the basic data.

Apart from the use of expression (1), there are two ways in which the adequacy of a trend-surface map can be assayed. First, the data density can be steadily increased until the resultant trend surface is stabilized—*i.e.*, is not significantly altered by inclusion of additional new data. Such a procedure may be impossible if data are unusually difficult to obtain (although it could be argued that the cost factor is unimportant if a correct result is sacrificed). If it were possible to compare trend components objectively one could compare maps for two independent variates, each based on one specimen per square mile, then two maps based on two specimens per square mile, then three specimens, and so on. When a comparable pair of maps is obtained, the data density will probably be adequate to give a reproducible result. Alternatively, if an objective measure can be made of a single trend component, the data density could be steadily stepped up until the desired confidence level is reached. Such procedures involve tests not available at present.

Second, when a suitable data density appears to have been achieved for one variable, it is possible to prepare trend surfaces for different

[5] Since this paper was prepared it has been found possible to use analysis of variance techniques to associate precise confidence levels with individual trend components (Dawson and Whitten, In press).

variates. If these new variates yield trends comparable with those of the first variate, the new results would appear to corroborate the original ones and thus suggest that each is a close approach to the complete trend of the sampled population, and a reasonable estimate of the target population represented by the sample. Such corroboration holds provided that the variables tested are independent rather than dependent. Dependence of data is a difficult problem and needs further exploration; the problem is most acute when results are based on percentage data.

In the present case $X_1 + X_2 + X_4 = 100$ per cent, and $X_5 + X_6 + X_7 + Y = X_2$ (where Y is part of the color index). Chayes (1948) concluded that the ". . . condition that the sum of the variates is a constant insures that major constituents of rocks . . . stated as modes . . . will generally show negative correlation" (p. 425), since the ". . . *a priori* chance that a given correlation coefficient will be negative is at least 7/8 or 87%, and this is probably an underestimate . . ." (p. 417). The writer (Whitten, 1959a, p. 844) pointed out previously that ". . . strike lines will necessarily be sensibly parallel for variables Z_1 and Z_2 expressed as percentages where $Z_1 + Z_2 = 100$ per cent for each sample, but parallelism is unnecessary for three variables such as $Z_1 + Z_2 + Z_3 = 100$ per cent . . ." Thus the near parallelism of isopleths may be, but is not necessarily, an inherent property of the numbers. If the near parallelism is not a function of dependence among the variables, the coincidence probably must be attributed to some specific geological cause. Chayes (1960) has critically investigated the nature of the correlations between components of three- and four-variable (and M-variable) closed tables and has demonstrated relationships which imply that extreme care is required in employing any argument of corroboration.

If any one of the variables X_1, X_2, X_3, X_5, X_6, or X_7 is independent of the others then the argument by corroboration can be employed. If the correlation coefficients between variables are determined, the rank of the matrix of correlation provides evidence on the independence of the variates (*cf.*, Anderson, 1958), However, it is by no means certain that variates like those involved in this study are amenable to analysis by such a precise mathematical artifice. This suggestion is based on the fact that, as described above, the values of X_{obs} comprise a considerable residual term ($X_{obs} - X_{comp}$) in addition to the trend component, so that the observations are not precise determinations.[6]

In summary, there is apparently no adequate mathematically rigorous means currently available for determining the data density necessary for computing the modal composition and variability of a granitic rock body. Empiricism and subject-matter evaluation are required to evaluate the results presented in this paper and similar data from other areas. The writer believes that the present results give a guide for planning future research, and that when considerably more quantitative data are available more specific tests of significance may be evolved. Clearly there is an urgent need for complete sampling and testing of sampling schemes at all levels.

SUBJECT-MATTER EVALUATION

The difficulty of appraising the value of results has already been emphasized, and it does not seem possible to avoid employing standards which are peculiar to each problem investigated. The results of experiments 040001-040003 inclusive (Bunbeg area) and 040011-040012 (Thorr district) illustrate the point. In

[6] The matrix of correlation based on all the data for five variates (X_1, X_2, X_3, X_5, and X_6) listed in Table 6 for Thorr district is:

$$\begin{bmatrix} 1.0000 & -0.2461 & -0.2371 & -0.3279 & -0.3599 \\ -0.2461 & 1.0000 & -0.2829 & 0.2880 & 0.3381 \\ -0.2371 & -0.2829 & 1.0000 & 0.0070 & -0.0517 \\ -0.3279 & 0.2880 & 0.0070 & 1.0000 & 0.1379 \\ -0.3599 & 0.3381 & -0.0517 & 0.1379 & 1.0000 \end{bmatrix}$$

X_4 (= total feldspar percentage) is omitted because $(X_1 + X_2) + X_4 = 100$ per cent. From this matrix the following may be developed:

$$\begin{bmatrix} 1.0000 & -0.2461 & -0.2371 & -0.3279 & -0.3599 \\ 0 & 0.9394 & -0.8664 & 0.2073 & 0.2495 \\ 0 & 0 & 0.6290 & 0.0046 & -0.0464 \\ 0 & 0 & 0 & 0.8458 & -0.0263 \\ 0 & 0 & 0 & 0 & 0.8010 \end{bmatrix}$$

so that the rank of the original matrix is five, which suggests that five linearly independent vectors exist in this set of correlations. For the Bunbeg area (Table 3, 040026) the matrix of correlation with respect to individual values of X_1, X_2, and X_3 is

$$\begin{bmatrix} 1.0000 & -0.4932 & 0.2034 \\ -0.4932 & 1.0000 & -0.6377 \\ 0.2034 & -0.6377 & 1.0000 \end{bmatrix}$$

Considered as a determinant, the solution of 0.4367 is obtained, and since this value $\neq 0$, it could be concluded that three linearly independent vectors exist in this set of correlations.

If some test of this type is not immediately available, empirical heuristic argument must be employed in the interim.

the former experiments the near parallelism of strike lines on the quartz percentage, color index, feldspar ratio, and total feldspar percentage partial-trend surfaces for averaged data is taken as evidence that each of the surfaces is a close approach to the respective complete trend of the target population. Again, the linear-plus

clusions drawn. A different aspect of this problem is illustrated by each of the two groups of experiments referred to above.

In the Bunbeg area the existence of strong correlation coefficients (either positive or negative) between the variables would reduce the degrees of freedom in Figure 4D, so that the

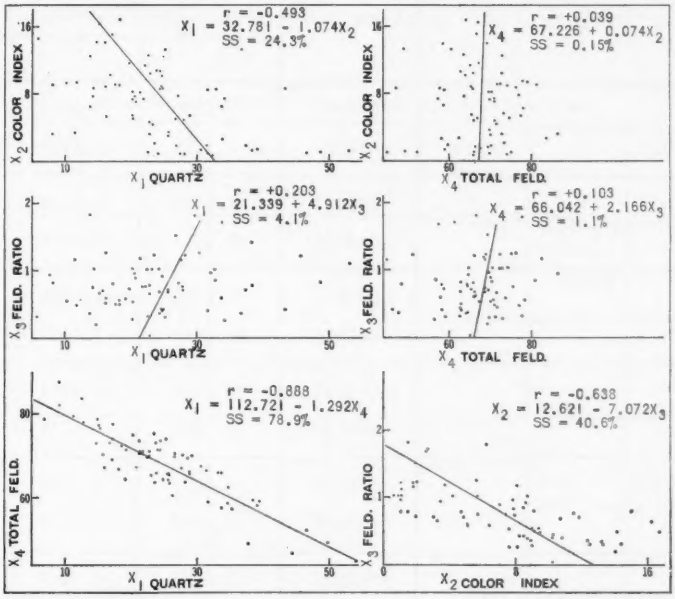

Figure 8. Least-squares linear regression lines for all data from Bunbeg area of the "older granite" of Donegal (experiment 040026). Each equation is that for the linear regression line. FELD. RATIO = Ratio of microcline to plagioclase; TOTAL FELD. = Sum of microcline and plagioclase; r = correlation coefficient; SS = percentage of sum of squares accounted for by the linear regression line.

quadratic partial-trend surfaces for both total feldspar and epidote percentages are sensibly similar (geometrically) to the surfaces for color index, feldspar ratio, and sphene percentage illustrated in Figure 6 (experiments 040011–040012). The geometrical similarity of these hyperboloid surfaces strongly suggests that each surface closely approximates the respective complete regional trend for each of the five variates within the target population.

Inferences such as these are based on subject-matter evaluation grounded in geological experience, which thus involves empiricism rather than proof, and in consequence it is difficult to set limits on the accuracy of the con-

presentation of four surfaces would give an unwarranted appearance of duplicated conformity.

In the granitoid rocks of Donegal it might be anticipated that negative correlations exist between quartz and both color index and total feldspar, and that a positive correlation might exist between color index and total feldspar. The results of regression analysis between the four variables for the 71 individual modes within the Bunbeg area are shown in Figure 8; although the negative correlations between quartz and both color index and total feldspar are well marked, the regression line for color index and total feldspar has very little sig-

nificance. The regression lines indicated in Figure 8 have been computed with the IBM 650 by the method of least squares (Krumbein, 1959b); the coefficients a and b for the linear regression lines $X_o = a + bX_1$ are tabulated in Table 3. Table 3 also gives the data for the regression lines corresponding to the average values employed in experiment 040002. When a least-squares method has been used to compute the regression lines, the confidence level associated with each line may be determined by analysis of variance techniques, provided that the variance is stable. When percentage data are involved, it is reasonable to assume that the condition of stable variance is met where the values range between about 30 and 70 per cent. Making such an assumption, Table 4 shows that, for the linear regression of total feldspar percentage with respect to quartz percentage, the confidence level is very high for both the individual (040026) and the averaged (040025) data.

The proportion of the sum of squares accounted for by the regression lines relating the averaged data (040002 and 040025) is much larger than that for the individual data (040026), except in the case of total feldspar/quartz. In general, as the number of data (N) is increased, the same level of confidence attaches to regression lines which account for a decreased proportion of the total sum of squares (for 040025 and 040026, N equals 21 and 71, respectively). Because of the nature of the data, it is inappropriate to stipulate the minimum percentage of the sum of squares which causes a significant level to be associated with each linear regression line.

The regressions between these components are to be correlated with the near parallelism of the linear partial-trend surfaces, and both

TABLE 3. LINEAR REGRESSION-LINE DATA FOR BUNBEG AREA

Variables	Experiment	Linear coefficients		SS*
		a	b	
Total feldspar/quartz	040025	167.0699	-2.0916	73.56
	040026	112.7210	-1.2918	78.93
Total feldspar/color index	040025	64.3139	0.4718	25.42
	040026	67.2257	0.0738	0.15
Total feldspar/feldspar ratio	040025	70.8420	-4.1065	13.87
	040026	66.0417	2.1657	1.06
Quartz/color index	040025	36.6800	-1.4711	76.81
	040026	32.7806	-1.0738	24.30
Quartz/feldspar ratio	040025	14.5218	13.7504	48.32
	040026	21.3392	4.9116	4.14
Color index/feldspar ratio	040025	14.6233	-9.6314	66.80
	040026	12.6210	-7.0722	40.64

* SS is percentage of total sum of squares accounted for by the linear regression line.
In experiment 040025 $N = 21$ (of 040002), and in 040026 $N = 71$; the data are averages and individual specimens respectively.

phenomena are expressions of inherent properties of the data. These relationships might be taken as detrimental to the theory that near parallelism of surfaces reflects the target-population trends. However, this would not appear to be the case. The regression lines indicated in Figure 8 merely confirm the mineralogical gradients seen in the field and demonstrate that parallel trend surfaces should be anticipated in this case; this is added confirmation that the trend-surface approach is an effective means of graphically portraying the mineralogical variability of a rock mass. With the surfaces evolved by experiments 040001 and 040003 such near parallelism does not obtain, so the modes employed are poor estimators of the target-population trends. However, with 040002 a sufficiently good sample has been taken to demonstrate on a map the independently established correlations, despite the local variability of each variable—it is this local variability that accounts for the vagaries (nonparallelism of strike) in the 040001 and 040003 linear surfaces. Hence, the positive and negative regressions between the mineralogical variables strengthen the argument that, for the granite studied, experiment 040002 provided a good estimate (in terms of a map) of the total trend of the target population. When two variates are not independent, the trend components of each do not provide completely independent and different information.

Some of the variables employed are better indicators than others of the complete trend, even when a limited number of specimens is available. Figure 9 shows that the three partial-trend surfaces for color index bear much more similarity than those for, say, total feldspar percentage. This suggests that one specimen per square mile gives a better estimate of the color-index trend in the target population than of the complete trend of total feldspar percentage in the target population. It is difficult to erect any precise methods by which to predict such differences; the proportion of the sum of squares accounted for by each surface will give only an indication of the importance of that surface in relation to the particular data used, and not necessarily in relation to the target population. Thus, the fact that the proportion of the sums of squares for each color-index surface is higher in Figure 9 than that for each of the three quartz-percentage surfaces, does not, in itself, imply that the 040001 color-index surface is a reasonable approximation to the trend component of the target population; the suggestion that this color-index surface is a reasonable approximation is based

TABLE 4. ANALYSIS OF VARIANCE FOR LINEAR REGRESSION OF TOTAL FELDSPAR PERCENTAGE (X_0) WITH RESPECT TO QUARTZ PERCENTAGE (X_3) IN BUNBEG AREA

Experiment	Source of variation	Sum of squares	Degrees of freedom	Mean square	F ratio	Confidence level
040025 N = 21	Due to linear regression	175.6156	1	175.6156	52.868	$\geq$ 99.95%
	Deviations from linear	63.1145	19	3.3218		
	$N.\sigma_{x_o}^2$ =	238.7301	20			
040026 N = 71	Due to linear regression	3724.2985	1	3724.2985	258.542	$\geq$ 99.95%
	Deviations from linear	933.9416	69	14.4050		
	$N.\sigma_{x_o}^2$ =	4718.2401	70			

on the similarity of all three of the color-index surfaces. It is difficult to be certain of the factors that cause one variable to yield better estimates than others. However, color index in these rocks seems better than some other variables because:

(1) The grain size of the mafic minerals (principally biotite and subordinate hornblende and iron ore) is smaller than that of the felsic constituents. Chayes (1956, p. 80) elaborated evidence which appears to establish that a longer traverse is needed to obtain a satisfactory mode of a coarser-grained rock. Probably if one variable consistently occurs as smaller grains in a given rock it may be estimated more accurately in a given traverse.

(2) On the microscale the mafic components may be distributed more homogeneously than the other phases such as quartz and microcline microperthite. In hand specimens of the "older granite" of Donegal (and many other granites) the felsic minerals appear to have a rather blotchy distribution; under such conditions standard point-counter measurements may not give equally good estimates for all the variables assayed in a suite of rocks.

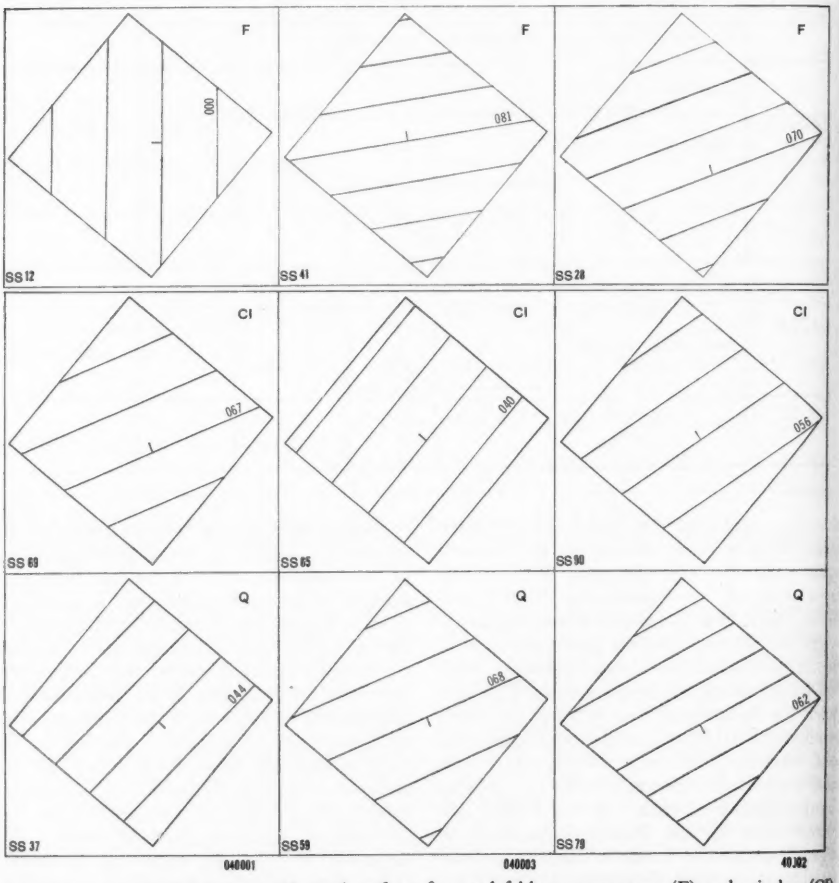

Figure 9. Comparative linear partial-trend surfaces for total feldspar-percentage (F), color-index (CI), and quartz-percentage (Q) data in experiments 040001 and 040003 (both with one specimen per square mile) and 040002 (averages of all specimens in each square mile). SS—percentage of sum of squares accounted for by the trend surface

The significance of regression relationships between components of granitoid rocks is being studied more closely. Reference to the Thorr district shows that even for one mass of granitoid rocks interrelationships between the partial-trend surfaces and the linear regressions between constituent minerals may vary from those recorded for the Bunbeg area. Thus, generalizations are difficult to enunciate at this stage. The similarity of the hyperboloid linear-plus quadratic partial-trend surfaces (Figs. 6, 7D) appears to be conclusive evidence that these surfaces are good estimates of the trend components of the target population; however, Figures 7A and B show that quartz and hornblende percentages yield surfaces disharmonious with those for the other variables. If the reasoning applied above to the Bunbeg area is correct, it could be assumed that either (a) the hyperboloid surfaces approximate the trend components of the target population, but that insufficient data are available for quartz and hornblende to counteract the effects of local variability about the complete trends of the sampled population, or (b) the correlations between the variables which obtain for the Bunbeg area do not occur with respect to the Thorr district (*e.g.*, that for the Thorr district data the linear regression line between quartz percentage and color index has little significance). In this case, the general similarity of the quartz and hornblende surfaces might indicate that the surfaces illustrated in Figure 7 are reasonable approximations for these variables in the target population.

Relationships between the variables in Thorr district contrast with those of the Bunbeg area and indicate some of the complexities of quantitative modal variation problems. These complexities involve some major difficulties which require further study.

Table 5. Proportion of Sums of Squares Accounted for by Different Partial-Trend Surfaces Based on Averaged Data for Bunbeg and Thorr

Variable	Bunbeg (040002)		Thorr (040012)	
	Linear	Linear plus Quadratic	Linear	Linear plus Quadratic
Quartz percentage	79.2	84.3	42.6	43.8
Color index	90.4	90.5	28.5	73.0
Feldspar ratio	64.8	68.3	36.5	67.1
Total feldspar percentage	28.2	53.4	39.9	47.1
Hornblende percentage	-	-	49.5	55.2
Sphene percentage	-	-	18.7	47.2
Epidote percentage	-	-	15.4	27.7

In the Bunbeg area the quadratic trend components add little to the linear components, and thus (especially for color index) the linear partial-trend surfaces are reasonable estimates of the complete trends. For Thorr district the situation is more complex. Table 5 shows the increased proportions of the total sums of squares accounted for by the linear-plus quadratic partial-trend surfaces for the two areas. In Figure 10 the linear partial-trend surfaces (averaged values) for Thorr are shown for comparison with Figures 6 and 7. The quadratic component adds little to the percentage of the

sums of squares accounted for by the linear partial-trend surface for quartz (Table 5), and little in the case of total feldspar and hornblende percentages; the three linear surfaces, involved are shown in Figures 10A, D, and E, and the strikes of these planes are subparallel. Although Figure 10 is based on average data the general lack of parallelism of the linear surfaces is striking. (*cf.* Figure 4, the near parallelism at Bunbeg.) The heterogeneity of these linear components contrasts with the similarity of structure demonstrated by the higher-degree

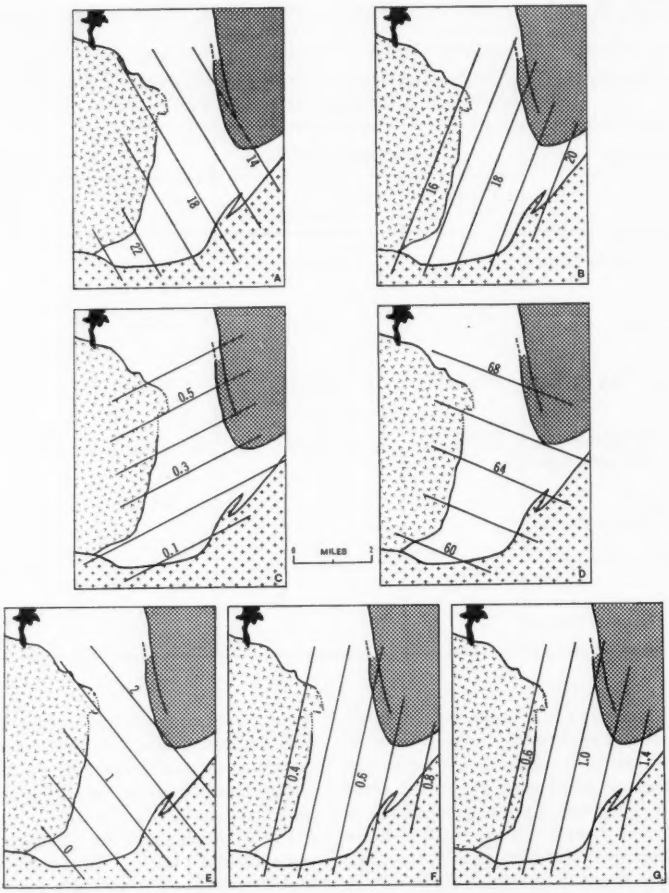

Figure 10. Linear partial-trend surfaces based upon averaged data for the "older granite" of Thorr district (experiment 040012). Dot pattern—envelope metasedimentary rocks; small v and cross patterns—younger granites. A. Quartz-percentage surface, which accounts for 42.6 per cent of the sum of squares. B. Color-index surface, which accounts for 28.5 per cent of the sum of squares. C. Microcline/plagioclase ratio surface, which accounts for 36.5 per cent of the sum of squares. D. Total feldspar-percentage surface, which accounts for 39.9 per cent of the sum of squares. E. Hornblende-percentage surface, which accounts for 49.5 per cent of the sum of squares. F. Sphene-percentage surface, which accounts for 18.7 per cent of the sum of squares. G. Epidote-percentage surface, which accounts for 15.4 per cent of the sum of squares

surfaces of Figures 6A, C, E, and 7D (and possibly 7C); these surfaces relate to variables for which the quadratic component adds significantly to the sum of squares—color index, feldspar ratio, sphene, and epidote percentages (and possibly total feldspar percentage). Figures 6, 7, and 10 and Table 5 suggest that the results from the Bunbeg area cannot be extrapolated directly in order to draw conclusions about Thorr district. The Thorr variates investigated fall into two groups:

(1) Quartz percentage, hornblende percentage, and probably total feldspar percentage have subparallel strike lines on the linear partial-trend surfaces (Figs. 10A, E, D, respectively) and little contribution from the quadratic component (as measured by reduction in the sum of squares). For these variables the linear appears to be a reasonable estimate of the complete trend. (*cf.* Bunbeg experience.)

(2) Color index, feldspar ratio, and sphene and epidote percentages show geometric similarity of the linear-plus quadratic partial-trend surfaces (Figs. 6A, C, E, 7D, respectively), and a considerable contribution from the quadratic component (as measured by reduction in the sum of squares). For these variables the linear-plus quadratic components appear to be reasonable estimates of the complete trend. Table 5 shows that a considerable proportion of the total sum of squares remains unaccounted for by the linear and quadratic components. The cubic components for color index and feldspar ratio add only 2.5 and 3.7 per cent, respectively, to the sum of squares. Thus, the cubic and all higher-degree terms are probably to be associated with the deviations rather than the complete trend.

It seems probable that groups (1) and (2) have features which cause them not to relate directly to assumption (a) or (b) erected above. Assumption (b) concerns regression relationships, and in Table 6 the linear regression-line data for the Thorr district variables (040012)

TABLE 6. COMPARATIVE LINEAR REGRESSION-LINE DATA FOR ALL DATA FROM THORR AND BUNBEG

Variables	Thorr district (N = 65)			Bunbeg area (N = 71)	
	a	b	SS*	b	SS*
Total feldspar / quartz	85.5755	-1.1851	73.19	-0.7741	78.93
Total feldspar / color index	178.5335	-6.4350	8.48	0.0738	0.15
Total feldspar / feldspar ratio	47.2553	53.1915	15.02	2.1657	1.06
Quartz / color index	26.6360	-0.4549	6.06	-1.0738	24.30
Quartz / feldspar ratio	19.9977	-4.8189	5.62	4.9116	4.14
Color index / feldspar ratio	18.8128	-3.1111	8.00	-7.0722	40.64
Sphene / quartz	1.1799	0.0294	12.95	-	-
Sphene / color index	-0.2770	0.0511	11.43	-	-
Sphene / feldspar ratio	0.6616	-0.0859	0.27	-	-
Sphene / total feldspar	-0.2600	0.0141	3.04	-	-
Sphene / hornblende	0.5623	0.0579	1.90	-	-
Quartz / hornblende	20.6312	-1.6841	10.75	-	-
Color index / hornblende	16.8480	0.8004	8.29	-	-
Feldspar ratio / hornblende	0.3058	0.0018	0.005	-	-
Total feldspar / hornblende	62.5201	0.8844	2.88	-	-

* SS equals 100 × (correlation coefficient)^2 or percentage of sum of squares accounted for by linear regression line.

are compared with those for averaged Bunbeg variables (040002). Although Table 6 refers to linear regression lines, most of the Thorr district variables (category 2 above) involve linear-plus quadratic partial-trend surfaces. Problems associated with the inter-relationships between linear and quadratic regression lines and linear-plus quadratic trend components are imperfectly understood and are being studied further. Each of the first six pairs of variates listed in Table 6, except total feldspar/quartz, involves at least one variable which requires the quadratic (or higher-degree) trend component to account for its areal variability (*i.e.*, complete trend). In consequence, it would be unwise to correlate regression and trend characteristics of Thorr district at present. However, there is not a direct relationship between the apparently analogous trend components and the correlation coefficents (relative to the linear regression lines). The correlation coefficients for the Thorr district variate pairs listed in Table 6 have been listed on an ordinal scale. In Table 7 the similar pairs of trend surfaces (Figs. 6B, D, F, 7A, B, C) are compared with this ordinal scale and the actual correlation coefficients. The implications of relationships such as these are the subject of further enquiry.

Despite the difficulties associated with regression analysis in relation to the Thorr district data, the assembled evidence seems to suggest that with four specimens per square mile the data density is adequate for reasonable approximations to the complete trend of each variable to be made. This conclusion is based mainly on the geometric similarity of the linear-plus quadratic partial-trend surfaces for group (2), and of the linear-trend surfaces for group (1). The dissimilarity of some trend surfaces to others is probably not evidence of an insufficient data density for this area but may indicate that trend components of higher degree are significant for some of the variates.

Table 7. Comparison of Correlation Coefficients of Variates Having Similar Trend Components in Thorr District

Figures showing surfaces	Variates	Rank in ordinal scale of SS from Table 6*	Actual correlation coefficient
6B and 6D	X_2 and X_3	8th	-0.283
6B and 6F	X_2 and X_6	4th	0.338
6D and 6F	X_3 and X_6	14th	-0.052
7A and 7B	X_1 and X_5	5th	-0.328
7A and 7C	X_1 and X_4	1st	-0.856
7B and 7C	X_4 and X_5	12th	0.170

* SS is percentage of sum of squares accounted for by linear regression line.

As studies are made of other areas further complicating factors will probably become apparent. Until experience is gained from several different suites of rocks, the quantitative variation characteristics of each area probably will have to be evaluated individually. Analyses of new data from both Donegal and some American localities have been initiated recently. In the *interim* it appears unwise to predict in advance the number of samples required to approximate the complete trend of any particular variable in a new sampled population (let alone the target population).

It is beyond the scope of the present paper to discuss the geochemical and petrogenetic significance of the analyses presented. However, the comparisons between the Bunbeg and Thorr data shown in Table 6 are of considerable interest; the confidence levels associated with the regression lines for the two areas are considerably different (other than for the total feldspar/quartz relationship). For example, in

the Bunbeg area biotite accounts for virtually the whole color index, and biotite and microcline-microperthite account for almost all the potassium. The high confidence level associated with the negative linear regression lines with respect to color index and the microcline/plagioclase ratio is important with regard to the distribution of potassium in the Bunberg area granites. In Thorr district this regression line also has a negative slope, but the confidence level is considerably reduced, although the positive regression line relating total feldspar to the feldspar ratio assumes some importance. Full evaluation of these and other relationships indicated by Table 6 must await the complete analysis of the "older granite" of Donegal.

CONCLUSIONS REGARDING REGIONAL VARIATION

In studying partial-trend surfaces for areal variation of various mineralogical parameters within the "older granite" of Donegal, average data, derived from four specimens within each 1-square-mile grid square, appear to yield reasonable results reflecting the consistent regional trends; such a data density may not be appropriate to other areas. By contrast, single specimens (for all variables) per square mile emphasize local inhomogeneity at the expense of the regional trend components.

Partial-trend surfaces based on average data are constructed more economically than comparable diagrams based on all the separate data points in each quarter square mile. These results will probably need some modification before generalizations applicable to any other plutonic rock body can be enunciated. However, with only slight modification it should be possible to extrapolate the conclusions, and the general method appears to yield the best means of investigating quantitative regional (areal) mineralogical variation.

SIMPLE RELATIONSHIPS IN DEVIATIONS FROM REGIONAL VARIATION

Deviation maps record the difference between a particular partial-trend surface and the observed datum at each specimen locality. Individual deviations in part represent variations from the complete trend of the sampled population. A more significant part of the deviation may represent the portion of the complete trend excluded from the surfaces of low degree (linear and linear-plus quadratic); in geologic terms this would indicate some local real departure of the mapped variable from the regional trend. As pointed out previously (Whitten, 1959a, p. 847), a consistent deviation structure necessitates reappraisal of the field evidence in a search for the geological explanation. It is important to evaluate the trends within a batholith to understand and appraise the bulk quantitative compositional variations of the mass, but evaluation of deviation structures may prove of greater importance in a petrogenetic study. The importance of deviations has been established in a number of geophysical studies (*e.g.*, Grant, 1957) and in the search for significant local variations in sedimentary rocks of interest to petroleum geologists. Where granitoid rocks have spatially replaced a varied series of metasedimentary rocks, remnants of which are preserved as ghost stratigraphy, local effects in the granite may be expected to show in the deviation pattern (Whitten, 1960) and may be a subject of study as important as the over-all trend components.

In order to study the sampling plan required to gain a quantitative picture of the deviation pattern, the deviations within the map area of Figure 6 may be evaluated. The feldspar ratio (Fig. 11A) and color-index (Fig. 11C) deviations from their respective linear-plus quadratic partial-trend surfaces are indicated in contoured maps. Figure 11A shows the positive deviations from the partial-trend surface based on each individual datum point within the quarter square-mile units (*cf.* Figure 6D); Figure 11C is a similar map based on deviations computed for data averaged within the square-mile grid unit (*cf.* Figure 6A). Figure 11B is a reconstruction of the ghost antiform composed of rafts of metasediment enclosed within the granitoid rocks. The petrogenetic relationship of the deviation patterns to the ghost stratigraphy has already been discussed (Whitten, 1960), and for the present purpose it will be assumed that the deviations have significance; it remains to explore the sampling necessary to define quantitatively the deviation pattern. The present deviation and partial-trend surface maps are based entirely on modal analyses of the granitoid rocks; the granitoid rocks contain metasedimentary relics, and these relics alone form the basis for the ghost stratigraphy indicated in Figure 11B. Not all deviation patterns in granitic rocks will relate to ghost stratigraphy and/or granitizational processes. A homogeneous magma might become differentiated or segregated by fluxion, gravity settling, vagaries of crystallization, or for a number of other

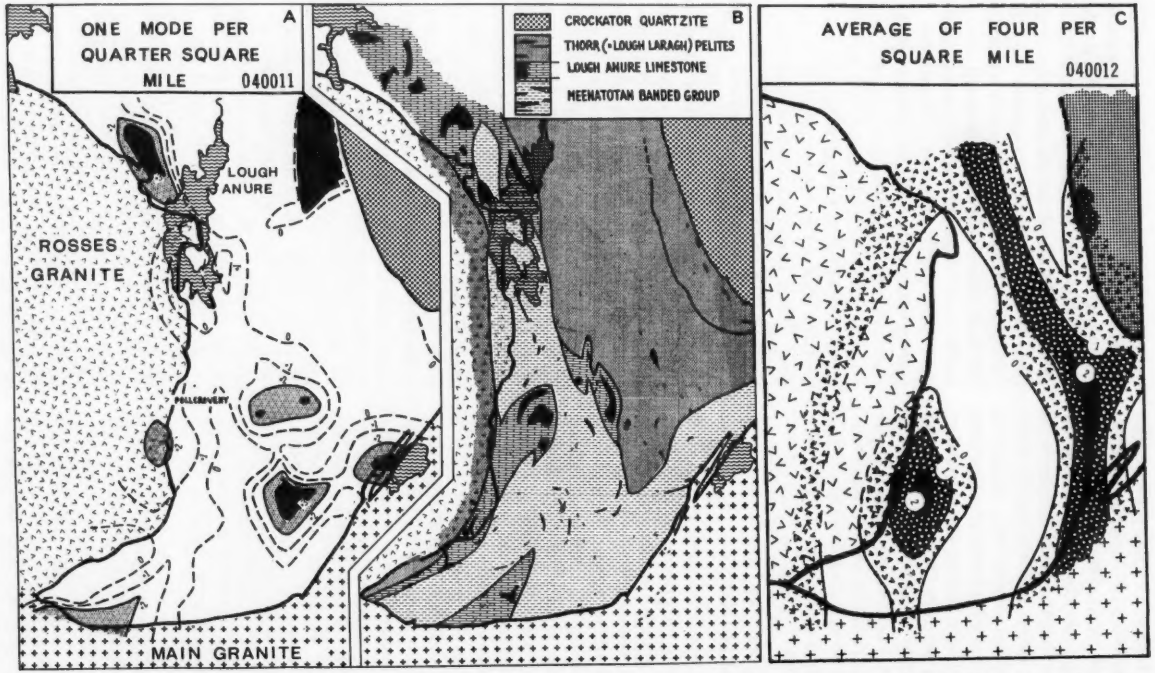

Figure 11. Comparison of positive deviation maps with the ghost stratigraphy within the "older granite" of Thorr district. Ripple lines indicate lakes or sea; rocks in whole figure are as identified in A or B. A. Positive deviation isopleths for microcline/plagioclase ratio (experiment 040011); deviations <0 omitted, 0–0.2 shown by isopleths, 0.2–0.3 shaded, >0.3 solid black. B. Ghost stratigraphy (based on Pitcher, 1953; Pitcher and Read, 1959). C. Positive deviation isopleths for color index (experiment 040012); deviations <0 omitted, 0–1 per cent shown by small v pattern, 1–2 per cent shown by white dot pattern, >2 per cent solid black.

reasons. Again, contamination by assimilation of wall rocks may cause magmatic inhomogeniety. Such local variations will be potential deviations from the regional trends of the sampled population of the intrusive rock.

Detailed examination of Figure 11 shows that the averaged color-index data (Figure 11C), although yielding a broad generalization of a northward-plunging antiform, do not permit specific correlation with the contaminating genetically significant metasedimentary enclosures within the granitoid rocks. By contrast, Figure 11A permits detailed correlation with Figure 11B; the high positive deviations correlate with the Lough Anure limestone horizon in the ghost stratigraphy. The main discrepancy occurs near Pollcrovehy, where, unfortunately, within 1 square mile, no granitoid rock specimens have been collected. Although different variables have been used in Figures 11A and C, the results are typical of those obtained for the seven variables examined. High microcline/plagioclase ratios in the granitoid rocks correlate spatially with zones of calcareous metasedimentary relics of the ghost stratigraphy; however, petrogenetic evaluation of these results will be deferred until data for the whole "older granite" are available.

For the present purpose maps based on individual observations are more useful than those based on averages. When correlations with local "disturbing" factors are sought, averaging the data will eliminate petrogenetically important local variability. Also, since it is desired to correlate local geologic influences with anomalous departures from the trends of the target population, the smaller the spatial interval between samples, the more critically the deviation pattern will be defined. In Thorr district specimens spaced more closely than one per quarter square mile would undoubtedly yield a more precise picture of the actual deviations from the complete trend. Although it is more costly of time and expense to use individual sample points within each quarter square mile (or even more closely spaced sample sites), this is necessary for deviation studies (but not necessarily for partial-trend surface studies) in an area comparable to Thorr district. These deviation studies point up the great need for similar studies in other areas, and only as a result of further work is it likely that optimum sampling schemes can be formulated. Consequently, in making an areal quantitative study of a batholith, the sample density needs to be varied according to the objectives of the project—the study of regional trends or of deviation patterns and correlations.

In deviation studies such as that described for Thorr district, it is particularly important to use objective sampling schemes. In some areas the factors which give rise to deviations may not be discerned in the field and thus are less likely to influence the specimen collector. In Thorr district the specimens were all collected before any idea of applying trend-surface analysis was conceived. If a geologist allows subjective factors to influence him in selecting specimens, the presence of visible rafts of metasediment might be a strong influence and a potential source of bias.

CONCLUSION

The aspects of quantitative mineralogical variation discussed above indicate the need for very careful analysis of the composition of a batholith (or similar large igneous mass) before broad petrogenetic conclusions can be drawn. Chayes and Fairbairn (1951, p. 711) state that ". . . we may conclude that with reasonable care and practice there is no reason why the reproducibility of quantitative thin-section analysis should not easily equal that of analytical methods ordinarily regarded as inherently far more precise . . .," and that in estimating the mineralogical composition of rocks a single thin-section analysis contains about as much, and about as reliable, information as a single chemical analysis (Chayes, *in* Fairbairn *et al.*, 1951, p. 66). The results given above, and Chayes and Fairbairn's claims indicate that, in assaying the composition of a batholith, a limited number of modes or chemical analyses could be, and probably will be, extremely misleading in relation to both the sampled population and the target population.

At the present time it is unknown how typical or atypical the different types of variation shown by the Donegal "older granite" are when compared with other granitoid complexes. Several projects are now on hand to obtain comparative data from a variety of different types of mass (*e.g.*, Dawson and Whitten, In press). In the interim extensive search of the published literature suggests that a large volume of new data will have to be collected if comparable results are to be obtained (Whitten, In press).

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Geology of the Seal Lake Area, Labrador

Abstract: The Seal Lake area covers about 800 square miles in central Labrador. Rocks are Precambrian and consist of the younger Seal Lake group resting unconformably on the Letitia Lake group and granites and anorthosite.

The Seal Lake group consists of metasedimentary rocks, basalt flows, and diabase sills totaling at least 34,200 feet in thickness. Shallow-water conditions prevailed during deposition of most of the succession; the uppermost beds, however, are of terrestrial origin. During sedimentation basalt was extruded over great areas, and diabase intruded the lower and middle and, to a lesser extent, the upper formations.

Pressures directed from south to north folded the rocks into an east-west–striking syncline. A granite mass was thrust northward over the Seal Lake sedimentary rocks, causing overturning of the syncline to the north. Considerable shearing and overturning took place along the contact between the granitic and sedimentary rocks, and the basal beds of the southern limb now dip south under the granite. Thrust faulting has removed considerable thicknesses of some formations of the southern limb.

The formations show a higher degree of metamorphism along the south side of the basin than along the northern limb because of the thrusting. The greenschist facies reached in the rocks of the Seal Lake group on the south limb represents the highest degree of metamorphism.

Most of the 250 small occurrences of copper minerals recorded in the area occur in two of the upper formations, the Salmon Lake and Adeline Island formations, directly associated with basaltic lavas or diabase sills or sedimentary rocks adjacent to them. The mineralization occurs in tension fractures or local shears which appear to be unrelated to the faults of the area. The minerals, in increasing order of abundance, are chalcopyrite, chalcocite, bornite, and native copper, and minor pyrite. Bornite, chalcopyrite, and pyrite are commonly associated with diabase sills, chalcocite and native copper with the flows. No copper deposits of economic importance have been found. The association of copper with only the upper flows and sills in such a thick sequence of mafic and sedimentary rocks indicates that only the final outpourings of magma were copper-rich.

Low-grade niobium–thorium–rare-earth deposits occur in the Letitia Lake group rocks in the southwest corner of the area.

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INTRODUCTION

More than 250 occurrences of copper mineralization occur in the Seal Lake area. Geological mapping, geophysical surveys, and diamond drilling of four of the better prospects has shown that these are not of economic value.

This paper describes the geology of the area, emphasizing the copper mineralization.

LOCATION AND ACCESSIBILITY OF AREA: The Seal Lake area is in central Labrador (Fig. 1). One can reach it in about a week by canoeing from the North West River up Grand Lake and then along the Naskaupi River into the east end of Seal Lake or by plane in 1 hour from either Goose Bay or North West River.

PREVIOUS WORK: Intensified exploration of the Seal Lake area began when R. A. Halet in 1946 supervised mapping and prospecting in the east-central part of the Seal Lake area for Dome Exploration (Canada) Limited.

During 1948 and 1949, H. S. Scott and H. K. Conn of the Photographic Survey Corporation of Toronto produced the first geological map of the area.

From 1950 to 1955, field parties of Frobisher Limited prospected in the area: E. L. Evans directed the work during 1950 and 1951, W. G. Robinson and B. W. Hester from 1952 to 1955. Operations included trenching, extensive geophysical surveys, and drilling of 31 shallow packsack drill holes, totaling 1448 feet.

During 1952, Christie, Roscoe, and Fahrig (1953) of the Geological Survey of Canada mapped the area.

During 1956, Kennco Explorations, (Canada) Limited optioned the ground from Frobisher Limited, and a party under the supervision of the senior author completed geological interpretation of photographs, reconnaissance and detailed geological mapping, extensive geophysical surveys, and geochemical surveys. The more promising copper prospects were investigated by 26 drill holes, totaling 7520 feet. Mann (1959, Ph.D. thesis, McGill Univ.) has recorded the data elsewhere.

Prior to the 1956 season, prospecting parties had located more than 250 small copper occurrences in the area. The problem remained as to whether the area contained an economic deposit or not.

Figure 1. Location of the Seal Lake area, Labrador

PRELIMINARY PHOTOGEOLOGICAL INTERPRETATION: Before the 1956 field season, a geological interpretation of photographs taken within the 925 square miles around Seal Lake was completed. This focused attention on the contact between the granitic and metasedimentary rocks, metasedimentary relicts in the granite, regional faults, fold structures, joint patterns, and drift-covered areas and permitted concentration of field work on critical areas.

The writers have compiled the maps presented herein from data collected during the 6 years prior to 1956 by members of the Frobisher parties and data collected during the 1956 field season.

ACKNOWLEDGMENTS: The writers acknowledge the assistance rendered by geologists of Frobisher Limited, especially Dr. W. G. Robinson and Mr. B. W. Hester.

All 46 members of the party performed their duties efficiently and capably.

Many other individuals and organizations rendered assistance during the summer, for which the writers express their appreciation.

PHYSIOGRAPHY

The Seal Lake area, like the rest of eastern Canada, has been extensively eroded by continental glaciers. In Labrador the ice radiated out from the northwest. In the Seal Lake district, the trend gradually swings eastward, and within the Seal Lake folded belt the ice movement was deflected to east-northeast along the strike of the metasedimentary rocks. The bottom of the ice sheet appears to have been tongued or grooved into underlying strike valleys along soft metasedimentary rocks between ridges of diabase. These valleys originated in preglacial times and had attained proportions comparable to those of the present landscape. The present shapes are typical of valleys eroded by continental glaciation.

TOPOGRAPHY AND DRAINAGE: The Seal Lake syncline can be identified by long, homoclinal ridges and gently curving strike valleys. Douglas and Drummond (1955) called the physiographic division the Naskaupi Mountains. The valleys contain a trellis drainage pattern and numerous elongated lakes, which indicate the structure of the overturned canoe-shaped Seal Lake syncline. Narrow ridges form the interstream divides, and their roughly accordant summits suggest peneplanation of the area, probably in the late Tertiary. From the north or south the surface appears extremely flat, but from the east or west the area is seen to be highly dissected with a local relief of 1000 feet or more. The area is generally slightly lower than the anorthosite massif to the north but higher than the drift-covered Red Wine River lowland to the south and west. Seal Lake is about 700 feet above sea level; the ridges average about 1800 feet above sea level. The most prominent peaks are Mount Pisa, Santa Claus Mountain, and Adeline Mountain (Wuchoo Mountain). To the north the Harp Lake anorthosite plateau rises well over 2000 feet above sea level.

The Thomas and Naskaupi rivers and their tributaries drain the area to the southeast.

PHYSIOGRAPHIC HISTORY: The Cenozoic physiographic history of the area, summarized, is as follows:

(1) Uplift during the late Tertiary resulted in youthful headward erosion of the rivers inland from the coast. The Naskaupi River and its tributaries eroded the softer metasedimentary rocks of the old peneplain, leaving long, parallel ridges which indicate the structure of the bedrock.

(2) During Pleistocene times the continental ice sheet, traveling from west to east, scoured out the preglacial strike valleys and probably deepened some. The surfaces of the intervalley ridges were polished and rounded.

(3) During ablation of the glacier, debris was washed into valleys, and dammed some, as at the east end of Pisa Lake. Outwash plains of coarse sand formed along the valleys of the Naskaupi and Thomas rivers. An esker follows the Naskaupi valley for the full length of the area. The melting of the glacier left the country mantled with erratic boulders and ground moraine up to 50 feet thick.

(4) In postglacial times the rivers cut into the outwash material. Some residual soil and extensive talus deposits have developed. Bedrock is exposed over only about 5 to 10 per cent of the area today.

GENERAL GEOLOGY

General Statement

The rocks in the area are the Precambrian Seal Lake group[1] and the underlying Letitia Lake group and granite-anorthosite complex (Fig. 2; Table 1).

The Seal Lake group, the most extensive, comprises arenaceous and argillaceous metasedimentary rocks interbedded with volcanic flows and intruded by diabase sills.

The group has been compressed into a major, east-trending, canoe-shaped syncline lying between an anorthosite massif in the north and a large granite mass to the south. This syncline has been overturned to the north (Fig. 3).

Within the overturned syncline, at least eight

[1] First named Seal Lake series by H. S. Scott and H. K. Conn (1950, unpub. report for Photographic Survey Corporation Ltd., Toronto). Name later changed to Seal Lake group by Kennco geologists

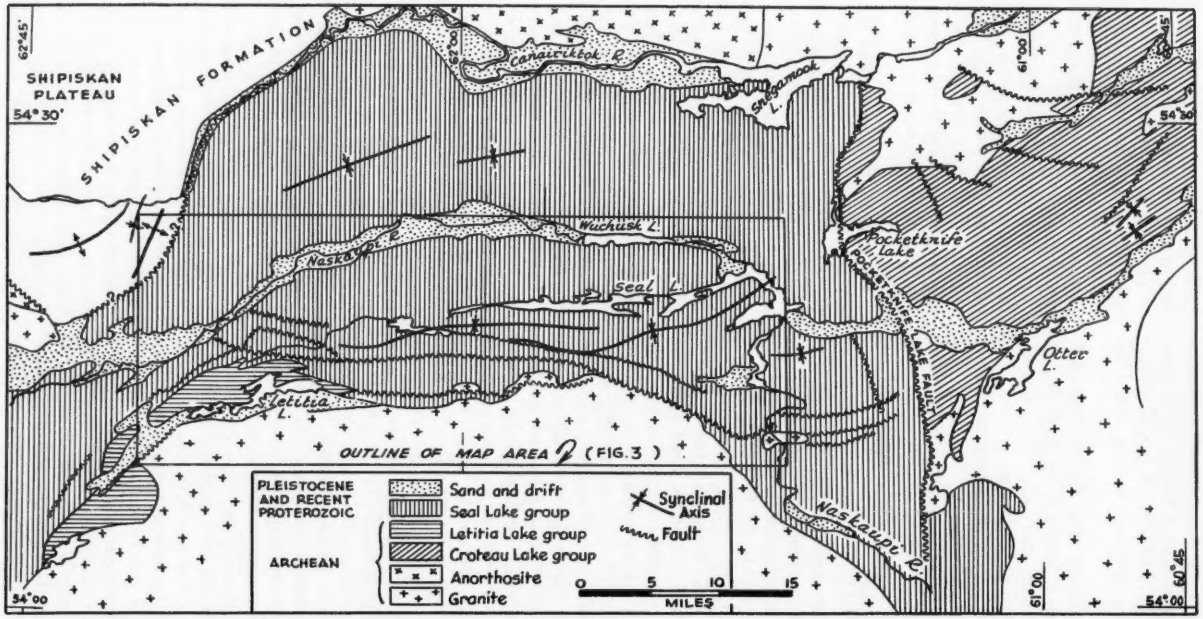

Figure 2. General geology of the area around Seal Lake, Labrador. Compiled by E. L. Mann

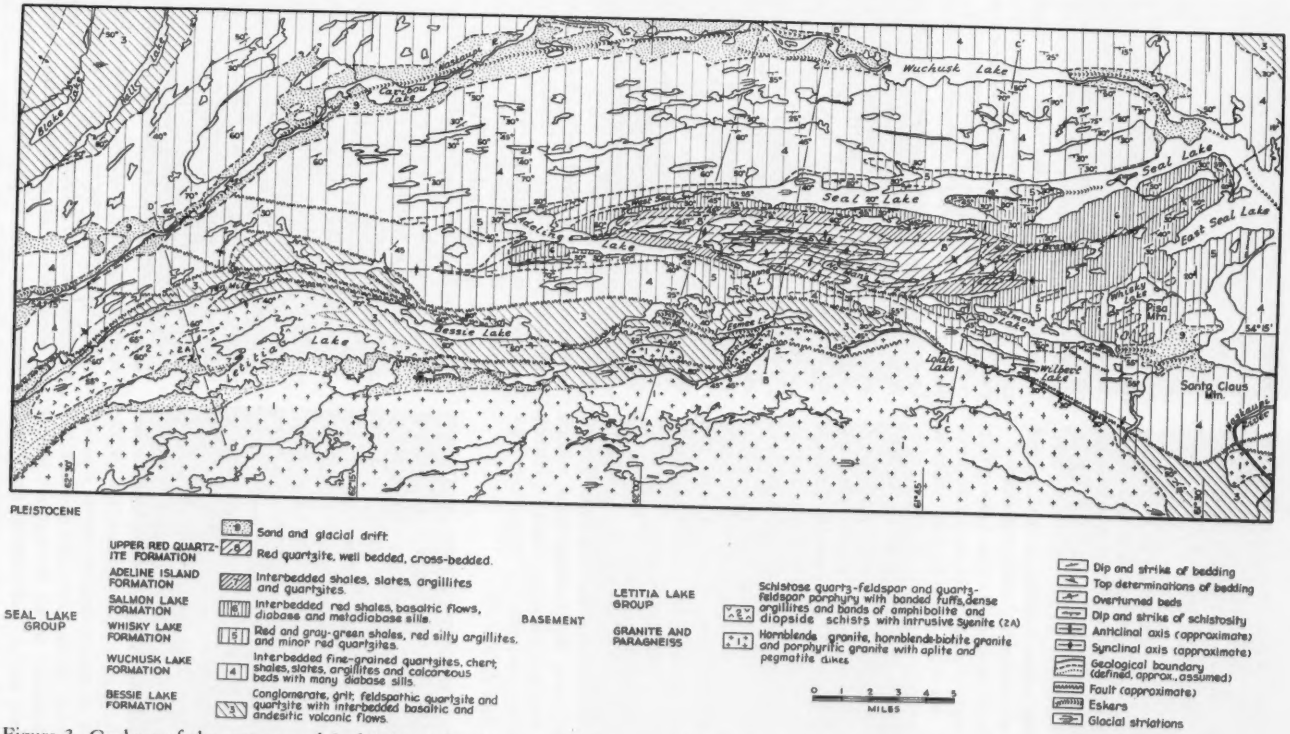
Wuchusk Lake
Seal Lake
East Seal Lake
Bessie Lake
Letitia Lake
Lolak Lake
Wilbert Lake
Santa Claus Mtn.
PLEISTOCENE
Sand and glacial drift.
UPPER RED QUARTZITE FORMATION
Red quartzite, well bedded, cross-bedded.
ADELINE ISLAND FORMATION
Interbedded shales, slates, argillites and quartzites.
SALMON LAKE FORMATION
Interbedded red shales, basaltic flows, diabase and metadiabase sills.
WHISKY LAKE FORMATION
Red and gray-green shales, red silty argillites, and minor red quartzites.
SEAL LAKE GROUP
WUCHUSK LAKE FORMATION
Interbedded fine-grained quartzites, chert, shales, slates, argillites and calcareous beds with many diabase sills.
BESSIE LAKE FORMATION
Conglomerate, grit, feldspathic quartzite and quartzite with interbedded basaltic and andesitic volcanic flows.
BASEMENT
LETITIA LAKE GROUP
Schistose quartz-feldspar and quartz-feldspar porphyry with banded tuffs, dense argillites and bands of amphibolite and diopside schists with intrusive syenite (2A)
GRANITE AND PARAGNEISS
Hornblende granite, hornblende-biotite granite and porphyritic granite with aplite and pegmatite dikes
0 1 2 3 4 5
MILES
Dip and strike of bedding
Top determinations of bedding
Overturned beds
Dip and strike of schistosity
Anticlinal axis (approximate)
Synclinal axis (approximate)
Geological boundary (defined, approx., assumed)
Fault (approximate)
Eskers
Glacial striations

Figure 3. Geology of the area around Seal Lake, Labrador. Data by Brummer, Mann, B. W. Hester, and members of Kennco field party (1956), and W. G. Robinson, B. W. Hester, E. L. Evans, W. Little, and members of Frobisher field party (1951–1955)

subsidiary fold axes parallel the major structure. The upper beds of the Seal Lake group and the outline of the contact of the granite buttress to the south show faint traces of north-south cross-folding.

The writers interpret shearing and mylonitization, together with marked linear features visible along the contact of the southern granite, as evidence of major thrust faults (Pl. 4, fig. 1). Slices of granite and Seal Lake rocks have been thrust over the upper formations with partial or complete elimination of some members.

The northeast-trending Letitia Lake group in the southwest corner of the area comprises interbedded, schistose, porphyritic volcanic and metasedimentary rocks, including quartz-feldspar porphyry, rhyolite, tuffaceous rhyolite, dense argillite, and minor agglomerate beds. The Letitia Lake group, too, has been overturned to the north and intruded by alkali syenite. These intrusions have given rise to rocks of complex mineralogy and low-grade niobium–thorium–rare-earth deposits.

South of the Seal Lake basin, granular hornblende granite, hornblende-biotite granite, and a porphyritic granite of the basement complex are exposed; all three have been cut by numerous aplite and pegmatite dikes.

Traces of copper mineralization are found throughout the rocks of the Seal Lake group, but the majority of the showings occur within a restricted stratigraphic range, mostly in mafic flows or in metasedimentary rocks near the contacts of thin diabase sills.

Basement Rocks

Granite and metasedimentary remnants. Granites surround the younger metasedimentary rocks. The largest mass occurs along the southern boundary of Figure 3. Thrusting of this large granitic buttress northward caused the deformation of the Seal Lake basin. The northern contact of the granite with the Seal Lake group has been followed for more than 25 miles west from the Naskaupi River to southeast of Bessie Lake where drift covers it.

Most of the granite exposures are of medium-grained hornblende granite with lesser amounts of hornblende-biotite granite and porphyritic granite. The granite rocks completely enclose small bands of quartz-sericite schist, chlorite-epidote schist, white quartzite, and epidote-amphibolite schist, which the authors believe are remnants of an older sedimentary succession that the granite has completely enclosed. Wherever these remnant bands occur, the contacts appear to be gradational through sheared and schistose granite to unaltered granite.

The trend of the foliation of the granite coincides in many places with the trend of the remnant sedimentary banding. The degree of foliation increases toward the northern fault contact with the younger metasedimentary rocks, suggesting that the foliation originated during the deformation of the Seal Lake basin.

Anorthosite. A large body of anorthosite and smaller bodies of granitic rocks east and west of it constitute the basement in the northern portion of the general area (Fig. 2). The contact between the Seal Lake syncline and the basement rocks follows the broad Canairiktok River valley, but recent glacial and alluvial sediments cover it. The first outcrops of anorthosite occur north of the Canairiktok River and form scattered, low hills. A high anorthosite massif rises abruptly to well over 2000 feet above sea level and extends north for many miles. This anorthosite covers a large area and extends well into the Nain district.

Letitia Lake group. The Letitia Lake group, first recognized and named by Mann south of Ten Mile Lake, occurs along the southwestern flank of the Seal Lake syncline and forms a narrow belt extending 25 miles southwest. The maximum outcrop width is 2½ miles, the estimated thickness 8500 feet. Drift obscures the relationship between the Letitia Lake group and the basement granites. The Bessie Lake formation, the basal formation of the Seal Lake group, lies unconformably on the Letitia Lake group. The unconformity is marked by a poorly developed, highly sheared, basal conglomerate which contains pebbles and fragments of the underlying rocks.

The Letitia Lake group includes: (1) a lower part composed dominantly of quartz-feldspar porphyry with lesser amounts of quartz porphyry and feldspar porphyry, and

(2) an upper complex succession of banded volcanic rocks, rhyolites, rhyolite porphyries, tuffaceous rhyolites, banded tuffs, fragmental rocks, argillites, and quartz-sericite schists.

The Letitia Lake group has been intruded by several plugs of Na_2O-rich syenite. The largest is 1½ miles long and one-third of a mile wide. A banded complex of aegirine-diopside amphibolite, and feldspathic schists is related to shear zones in these syenites. Niobium, thorium, and rare-earth minerals are associated with these schists and shear zones which occur near the gradational contact between the

Figure 1. Banded light quartz-sericite schist and dark chlorite-epidote schist occurring as remnants of sedimentary rocks in the southern granite massif, about 2 miles south of the contact with the Seal Lake group

Figure 2. Outcrop of massive to slightly foliated hornblende granite

Figure 3. Photomicrograph of hornblende-biotite granite showing slight trace of shearing. From location 2¼ miles south of contact between granite and Seal Lake group contact. 25×. Crossed nicols

Figure 4. Photomicrograph of mylonitized hornblende granite showing a broken feldspar crystal in a medium stage of cataclastic deformation. 25×. Crossed nicols

Figure 5. Outcrop of brecciated aegirine syenite containing fragments of syenite in dark gneissic bands of radioactive amphibolite. From Mann No. 1 prospect

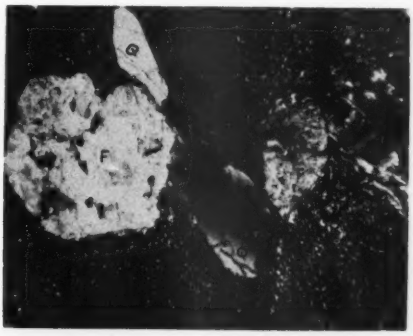

Figure 6. Photomicrograph of massive, quartz-feldspar porphyry containing euhedral crystals of microperthite (F) and phenocrysts of quartz (Q), in a matrix of microfelsite. Northern peninsula on Ten Mile Lake. 25×. Crossed nicols

SOUTHERN GRANITE MASSIF AND LETITIA LAKE GROUP, SEAL LAKE AREA, LABRADOR

BRUMMER AND MANN, PLATE 1
Geological Society of America Bulletin, volume 72

Figure 1. White Bessie Lake quartzites showing overturned cross-bedding. Top determinations indicated that southward-dipping beds on the south limb of the syncline have been overfolded. Exposure near Esmee Lake

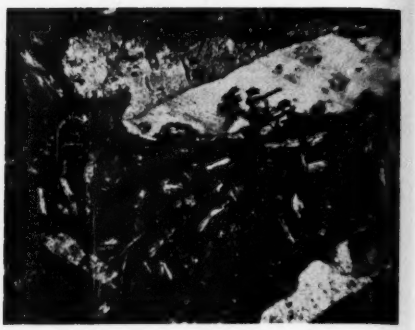

Figure 2. Photomicrograph of amygdaloidal basalt with plagioclase phenocrysts in a matrix of plagioclase, magnetite, and augite. Specimen from Bessie Lake formation on north limb of syncline and south of Canairiktok river. 25×. Crossed nicols

Figure 3. Outcrop of amygdaloidal basalt. Banding due to variation in abundance of amygdules and alteration along flow tops. From Salmon Lake formation south of Seal Lake

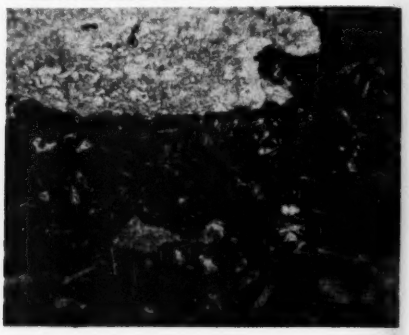

Figure 4. Photomicrograph of amygdaloidal basalt showing a large amygdule filled with calcite and quartz in a matrix of altered plagioclase and iron ore. From Salmon Lake formation south of Seal Lake on the Big Bight Peninsula. 25×. Plain light

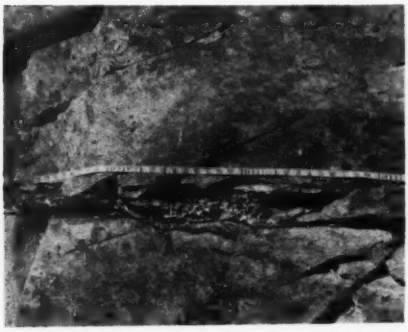

Figure 5. Quartz vein near base of diabase sill carrying irregular masses of native copper. Mineral occurrence 68 at west end of East Seal Lake in Salmon Lake formation. Inch scale

Figure 6. Veins cutting across diabase sill near lower contact with slates. Veins contain quartz, chlorite, and chalcopyrite. No. 1 zone, North Whisky Lake prospect, Salmon Lake formation

BESSIE LAKE AND SALMON LAKE FORMATIONS, SEAL LAKE AREA, LABRADOR

Figure 1. Photomicrograph of fine-grained saussuritized diabase containing a phenocryst of altered labradorite in a matrix of plagioclase and magnetite. From Wuchusk Lake formation, north of Seal Lake. 25×. Plain light

Figure 2. Photomicrograph of medium-grained diabase showing typical ophitic texture. Composed of laths of saussuritized feldspar (dark) enclosed in a large crystal of augite. From Wuchusk Lake formation south of Naskaupi River. 25×. Crossed nicols

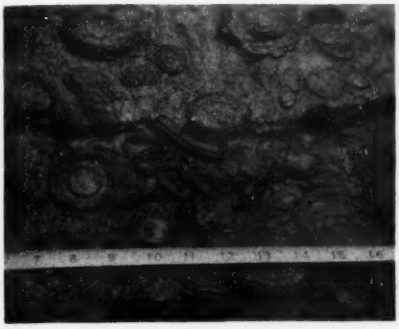

Figure 3. Algal limestone outcrop showing a broken fragment and several complete algal structures. Exposure about 2 miles north of the Naskaupi River. Inch scale

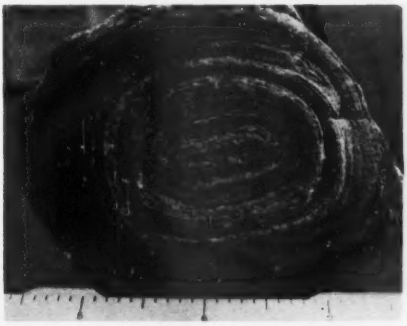

Figure 4. Single algal structure showing details of growth lines. Inch scale

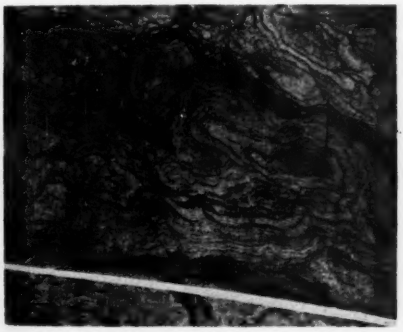

Figure 5. Deformed (?) algal structure in a sheared limestone. Length of photographed area = 48 inches.

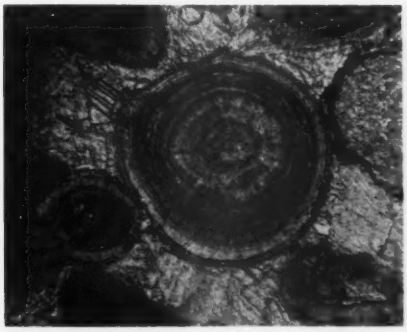

Figure 6. Photomicrograph of oölitic limestone from the Naskaupi River valley north of Ten Mile Lake. Oölite shows growth lines of carbonaceous material and calcite set in a matrix of calcite with small amounts of iron oxide and quartz. 25×. Plain light

WUCHUSK LAKE FORMATION, SEAL LAKE AREA, LABRADOR

Figure 1. Aerial photograph of area around Anne Lake showing Upper Red Quartzite formation (top), Whisky Lake formation (center), Bessie Lake formation (bottom). Upper Red Quartzite rocks show excellent synclinal fold structures. A thrust fault exists along prominent diagonal. Along this fault, some of Salmon Lake formation and most of Adeline Island formation have been faulted out of the area. Another major thrust is postulated along the cliff face south of Anne Lake, which has eliminated much of the Wuchusk Lake diabase (lower right) from the area. Along this fault, shales and argillites of Whisky Lake formation abut against greenschists of Bessie Lake formation. Photographed by Photographic Survey Corp. Ltd., Toronto (No. 238–5406–1533–22)

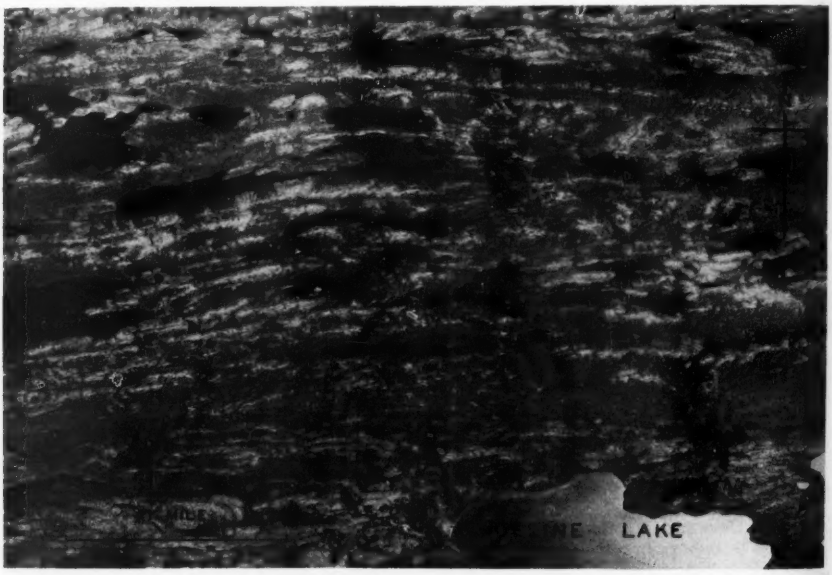

Figure 2. Aerial photograph of area north of Adeline Lake showing prominent homoclinal diabase ridges of Wuchusk Lake formation. Intervening valleys contain elongated lakes, are underlain by narrow bands of chert, quartzites, and shales dipping southward. Photographed by Photographic Survey Corp. Ltd., Toronto (No. 238–5407–1533–18)

AERIAL PHOTOGRAPHS SHOWING STRUCTURES AND DIABASE SILLS, SEAL LAKE AREA, LABRADOR

BRUMMER AND MANN, PLATE 4
Geological Society of America Bulletin, volume 72

porphyries and banded volcanic complex.

Structures within the group are complex and not readily defined because of the extensive cover of drift. The beds are overturned and dip 45° to 70° SE.

The rocks are generally schistose, but in a few places the porphyries tend to be massive.

Quartz veins occur throughout the porphyries; most are only a few inches wide, but a few are more than 10 feet long. Many fill tension joints, and a few larger veins follow the strike or cut obliquely across the foliation. A few have small amounts of pyrite, sphalerite, galena, and chalcopyrite.

South of Ten Mile Lake, lenticular amygdules in rhyolite flow tops contain quartz, calcite, and fluorite and minor galena, sphalerite, pyrite, and chalcopyrite.

TABLE 1. FORMATIONS OF SEAL LAKE AREA, LABRADOR

Age	Formation		Lithology
Pleistocene and Recent			Sand and glacial drift
			Unconformity
		Upper Red Quartzite formation (+2600′)	Red quartzites becoming argillaceous at the base, with some scattered pebbly bands
		Adeline Island formation (1400′)	Greenish-gray, black, red to purple slates and shales, pink and white quartzites, and argillaceous quartzite
		Salmon Lake formation (3000′)	Interbedded red, maroon, purple-red shales with minor green and gray-green shales, and abundant amygdaloidal basalts and some diabase sills
		Whisky Lake formation (3000′)	Red shales, gray-green slates and phyllites, red silty argillites, and minor red quartzites
Proterozoic	Seal Lake group (+34200 feet)	Wuchusk Lake formation (20000′)	Interbedded chert, variegated shales, phyllites, hornfels, argillites, fine-grained quartzites, and thin calcareous beds intruded by thick sills of normal diabase, metadiabase, and very coarse-grained diabase
		Bessie Lake formation (4200′)	Blue, white, and pink, clean, massive quartzites becoming feldspathic, gritty, and conglomeratic at the base, interbedded with vesicular and amygdaloidal basalts which have been metamorphosed in the south to chlorite-epidote schists. Equivalent to the Majoqua Lake formation (12500′) on the northern limb of the syncline
			Unconformity
	Letitia Lake group (+8500 feet)		Schistose quartz-feldspar porphyry, rhyolite, tuffaceous rhyolite, banded tuffs, argillite, and minor fragmental bands, intruded by aegirine syenite. Shear zones of banded feldspar, diopside, and amphibolite schists. May be equivalent to Croteau Lake group to the east
			Relationship Unknown
Archean	Anorthosite		Medium-grained to very coarse-grained anorthosite intruded by a few diabase dikes. Alluvium obscures the northern contact between anorthosite and the Seal Lake group.
	Granite and granitic gneiss		Hornblende granite, porphyritic granite, and hornblende-biotite granite with minor relict bands of quartz-sericite schist and chlorite-epidote schist. Cut by aplite and pegmatite dikes and quartz veins.

The lead-zinc-copper-niobium-thorium mineralization in the Letitia Lake group contrasts strongly with the dominantly copper mineralization in the mafic rocks of the overlying Seal Lake group.

The Letitia Lake rocks may extend eastward

beneath the younger Seal Lake rocks to reappear 40 miles away east of the Pocketknife Lake fault as the upper members of the Croteau Lake group (Fig. 2). The following features suggest this correlation:

(1) Both appear to be of pre-Seal Lake age.
(2) Both groups contain porphyries.
(3) Both groups contain radioactive minerals; none are found in the Seal Lake group.

Seal Lake Group

General statement. The rocks of the Seal Lake group occupy a broad synclinal basin which has been overfolded into an arcuate synclinorium at least 75 miles long and 28 miles wide. The dominant strike of the fold in the west is southwest; in the east the strike swings northeast and continues on toward the coast. The syncline contains at least eight major drag folds which parallel the major fold axis. The synclinorium is bounded on the north and south by anorthosite and granite of the basement complex and on the east is truncated by the north-south–trending Pocketknife Lake fault (Fig. 2). Along this fault rocks of the Seal Lake group abut against the Croteau Lake group. To the west the Seal Lake syncline narrows considerably before disappearing beneath glacial drift. To the northwest the broad alluvial and glacial outwash plain of the Canairiktok River valley separates the folded rocks of the Seal Lake basin from the relatively flat-lying drift-covered sedimentary rocks of the Shipiskan Plateau. These presumably are of Seal Lake age.

The Seal Lake group includes metasedimentary rocks, volcanic flows, and intrusive diabase and totals of at least 34,200 feet in thickness. Red and white quartzite and red slates and argillite predominate; graywacke, conglomerate, and calcareous rocks are present. Sills of diabase and metadiabase occur in parts of the Wuchusk Lake and Salmon Lake formations; basalt flows occur within the Bessie Lake formation and in the upper argillites and shales of the Salmon Lake formation. The rocks in the upper formations are predominantly red, suggesting shallow-water conditions of sedimentation.

Copper mineralization in the form of native copper, chalcocite, and bornite-chalcopyrite generally occurs near the contact of the volcanic flows and the red argillaceous metasedimentary members.

Bessie Lake formation. The Bessie Lake formation (Table 1) crops out along the southern margin of the Seal Lake syncline.

Toward the base of the formation the rocks become very feldspathic, and gritty beds are common along the unconformable contact with the older basement granite. These give way to two small lenses of basal conglomerate, exposed 2 miles apart south of Salmon Lake.

The larger, along the shore of Wilbert Lake, can be followed sporadically for more than 1000 feet. The pebbles of this conglomerate, which is exposed 10 feet from the granite, show marked elongation and flattening parallel to the schistosity, which dips 30°S. They average about 2 inches in length but are more than 6 inches long in places. The elongation ratio averages at least 6:1. The pebbles are predominantly fine-grained, granular, cherty quartzite and blue and white quartz, with scattered pebbles of jasper, greenstone, and shale. The matrix contains abundant pink feldspar which was probably derived from granite, but no granitic pebbles have been noted, probably because granite boulders were scarce during sedimentation. The arkosic matrix is very fresh, and thin sections show only small amounts of sericite. Magnetite octahedra are scattered through the matrix. In the granite to the south, disseminated pyrite is abundant. Abundant veinlets of fine-grained specular hematite and limonite are common.

The second exposure of basal conglomerate is on the southeast shore of Lolah Lake. Lithologically, the rock is almost identical with the conglomerate south of Wilbert Lake. Here, conglomerate occurs directly over schistose granite, proving that the Seal Lake group is younger than the granite. The granite south of the contact is essentially fresh and unaltered except in a narrow very schistose zone. Slight mylonitization is present, especially in the immediate vicinity of the contact, and the ferromagnesian minerals are completely chloritized. The observations indicate that shearing took place during folding, along this contact.

The Bessie Lake formation varies considerably in exposed width. South of Salmon Lake, where faulting has removed most of the upper members, it is only a few hundred feet; east of Bessie Lake the formation reaches its maximum width of more than 3 miles, possibly because of duplication by thrusting. The formation can be traced almost continuously for more than 20 miles southwest of the map area (Pl. 4, fig. 1).

Where the Bessie Lake formation is best developed, at least half the thickness is composed of thin chlorite-epidote schist bands representing metamorphosed basaltic flows, intercalated in the quartzites. Between Ten Mile Lake and Esmee Lake, basaltic flows are represented by a thick succession of greenstone schists.

A similar but less metamorphosed succession constitutes the base of the Seal Lake group

along the northern limb of the syncline. Previous geologists have named it the Majoqua Lake formation; the present writers consider it to be the northern equivalent of the Bessie Lake formation. Heavy drift prevents continuous tracing of the rocks around the western end of the syncline, but the formations appear to be continuous around the eastern end of the syncline along the Pocketknife Lake fault.

The deposition of coarse-grained arenaceous sediments continued after the end of the periods of volcanic outpourings, but for mapping purposes the Bessie Lake formation is demarcated by the termination of volcanic activity in the succession.

Wuchusk Lake formation. The Wuchusk Lake formation (Table 1) overlies the Bessie Lake. The base of the formation is the lowest diabase sill.

The outcrop area is typified by prominent parallel cuestas or homoclinal ridges of diabase (Pl. 4, fig. 2), which persist for miles. Long, narrow, strike valleys between the ridges generally follow thin bands of less resistant interbedded metasedimentary rocks.

CALCAREOUS ROCKS: Calcareous rocks form a very minor portion of the succession, but calcareous partings are common in the shales, cherts, and quartzites.

Algal structures (Pl. 3, fig. 3) are exposed in a 30-foot band of fine- to medium-grained, gray limestone along a low ridge about 2 miles north of the Naskaupi River and 3½ miles northwest of Wuchusk Lake. These are "stromatolites" of the general type, *Collenia* Walcott (D. J. McLaren, 1956, unpub. rept. No. PcI McL-56/57, Geol. Survey Canada).

Another ridge of limestone in the Naskaupi River valley north of Ten Mile Lake contains a zone of deformed algal structures and a zone of fine-grained oölitic limestone (Pl. 3, fig. 5).

DIABASE: The diabase and metadiabase sills of the Wuchusk formation range from fine-grained at the chilled margins to very coarse in the center of the sills but are remarkably consistent in composition throughout the area. Field evidence suggests that the diabase sills are completely concordant with the adjacent sediments.

The sills average about 800 feet in thickness, but geological sections of the area indicate that they range from about 100 to more than 2000 feet.

The diabase shows typical ophitic texture (Pl. 3, fig. 2). It consists of plagioclase laths set in large crystals of pigeonite. Smaller grains are commonly poikilitically included in the pyroxene. Fairly large, rounded spots, consisting mainly of serpentine, indicate the earlier presence of olivine. The plagioclase is extensively saussuritized, but unaltered remnants reveal an original composition of labradorite. Chlorite is abundant as a deuteric alteration product; apatite, ilmenite, pyrite, and chalcopyrite are accessory minerals.

The diabase along the northern limb of the syncline shows little trace of metamorphism other than saussuritization of the plagioclase. Metamorphism is more pronounced along the southern limb, and most of the original minerals have been affected. The altered rock contains remnants of an ophitic texture, but the plagioclase is highly saussuritized and rich in Na_2O. Very little of the original augite remains: it has been altered to chlorite and epidote. These rocks of the southern limb of the syncline are best termed "metadiabase." A few occurrences on the northern limb of the syncline probably also deserve the name, but in general the rocks there still appear relatively unmetamorphosed.

The diabasic sills constitute almost two-thirds of the Wuchusk Lake formation, but nowhere in the area have any crosscutting relationships been observed. These sills have been folded and deformed to the same degree as the metasedimentary rocks.

Whisky Lake formation. The rocks of the Whisky Lake formation (Table 1) are predominantly argillaceous. They can be traced continuously around the Seal Lake syncline. The formation is distinguished from the underlying Wuchusk Lake formation by a broad belt of red shales and argillites and the absence of sills.

Salmon Lake formation. Except for the presence of widespread basaltic flows and some metadiabase sills intercalated with the shales, the Salmon Lake formation is very similar to the Whisky Lake formation (Table 1). The contact is marked by the first lava flows in the succession of red shales; the contact of the Salmon Lake formation and the younger Adeline Island formation is marked by the final disappearance of the lavas from the succession.

Traces of copper minerals are abundant in the Salmon Lake formation. Occurrences are all small and of no economic value.

SHALES AND ARGILLACEOUS ROCKS: No red argillite occurs in the Salmon Lake sequence. Only weak traces of original bedding remain, and these are generally masked by flow cleavage, the result of folding. Pronounced crenulation and puckering of the argillaceous material

is common in some areas. Fine partings, lenses, and nodules of calcareous material are common in the red shales. The rocks are cut by quartz-filled tension fractures and veins of quartz and calcite which contain minor amounts of copper sulfides and malachite. Native copper, chalcocite, bornite, and secondary malachite in these veins appear to be associated with the basaltic flows in the vicinity.

Thin-section studies indicate that the red shales are composed of a turbid mass of very fine-grained clay minerals and rock flour with abundant hematite. The hematite indicates that oxidizing conditions existed at the time of the deposition of the red shales. None of the iron oxides shows recrystallization into specularite; this indicates that the regional metamorphism was of low grade.

A few bands of well laminated green and gray-green shales are interbedded with the red shales of the Salmon Lake formation. These bands are similar to the green shales and slates of the Whisky Lake formation.

The gray-green color of some shale bands indicates that reducing conditions existed during sedimentation.

QUARTZITES AND ARENACEOUS ROCKS: A few very thin bands of quartzite near the top of the Salmon Lake formation, found only in a few places along the south shore of Seal Lake, grade into gritty shales composed of scattered quartz grains in an argillaceous matrix. Colors vary from red and pink to gray and white bands. Thin partings of sericitic phyllite and bands of argillaceous material are common.

AMYGDALOIDAL BASALT: Amygdaloidal basalt, the most characteristic rock type (Pl. 2, figs. 3, 4) of the Salmon Lake formation, occurs intercalated with red shales and associated metasedimentary rocks. It is well exposed from Salmon Lake to the east end of Seal Lake; along the southern shore of Seal Lake to Adeline Lake; and along the southern shore of Adeline Lake to Anne Lake.

The amygdaloidal basalts are gray green; fresh surfaces commonly show a reddish or purple mottling. Flow textures are rare, but do occur. The rocks are sheared, and alteration is common. Most of the flows contain bands of amygdules filled with calcite and small amounts of quartz. Flow textures are rare. Flow banding is clearly exposed immediately south of the western end of Seal Lake: the individual flows range from about 3 to 5 feet in thickness and contain abundant amygdules filled with quartz and calcite along the edges of the flows. A marked red zone near the contact and also surrounding some of the amygdules is absent from the most massive portions of the flows.

The senior author believes that the flows erupted subaerially because pillow structures are absent and because boundaries between the flows are oxidized.

Successions of individual flows range from 50 feet to 200 feet in thickness; red shales and slates separate these successions. The number of successions of amygdaloidal basalt in the formation varies, but four or five are commonly found. A small basaltic dike along the southern shore of Seal Lake near Big Bight cuts across the red shales between two successions of basaltic flow rocks and is believed by the authors to be a feeder channel for the outpourings of volcanic rocks on the surface. This represents the only known example of crosscutting relationships within the Seal Lake group. All other volcanic and intrusive rocks are conformable with the sedimentary rocks in the basin.

Columnar jointing has been observed in a few places along the cliffs of the southern shore of Seal Lake. Generally the amygdaloidal basalts are not well jointed, but small tension cracks are common. Many of these are filled with white vein quartz and minor calcite or to a lesser extent with hematite, and a fibrous amphibole.

The quartz-calcite veins commonly carry small amounts of native copper and chalcocite. Veins of quartz, epidote, chlorite, and albite contain small amounts of chalcocite, bornite, chalcopyrite, pyrite, and molybdenite. More than 250 occurrences of copper minerals have been discovered in the area. The majority are in the Salmon Lake formation and are minor, consisting of slight malachite staining on the surface and specks of chalcopyrite in the basalt flows. A few show copper mineralization over appreciable lengths and widths.

Detailed investigation of a few of the more promising showings indicated that the copper mineralization is sporadic and of no economic consequence. Native copper occurs mainly in the basalt flows or in the red shales in contact with them in tabular to platy masses in small fractures filled with quartz and calcite. Specimens of high grade do not occur in sufficient numbers to warrant development. The copper appears to have originated from the amygdaloidal basalt or small diabase sills in the area. Semiquantitative spectrographic analyses of the diabase and lava, however, give only trace a

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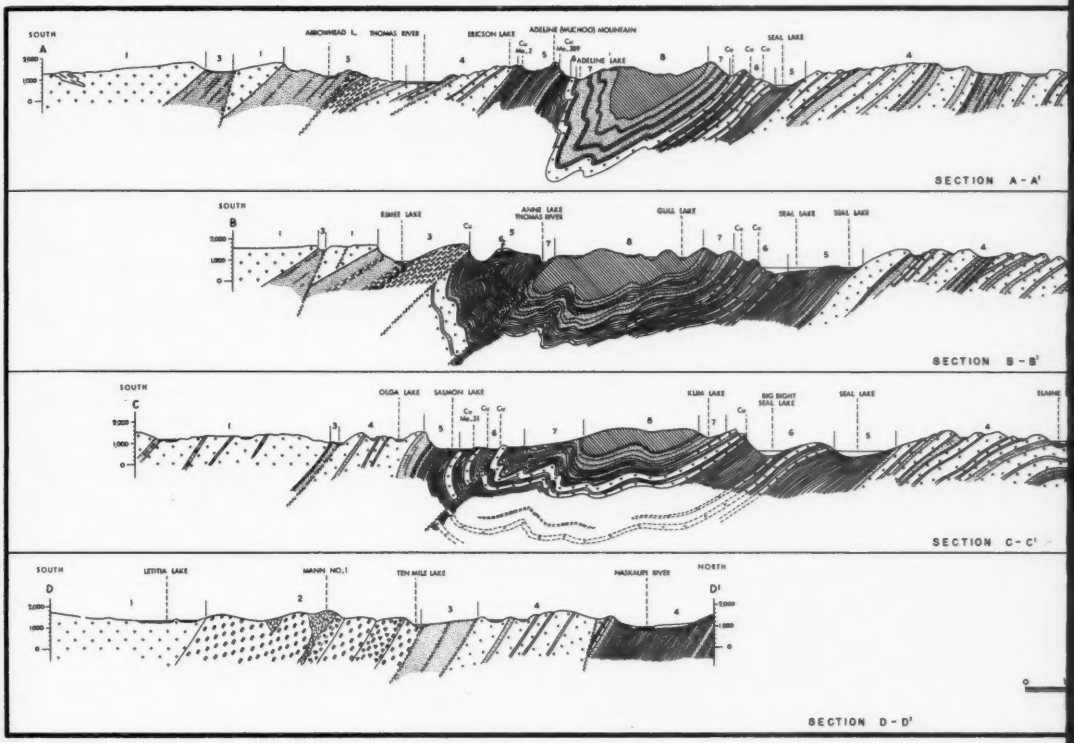
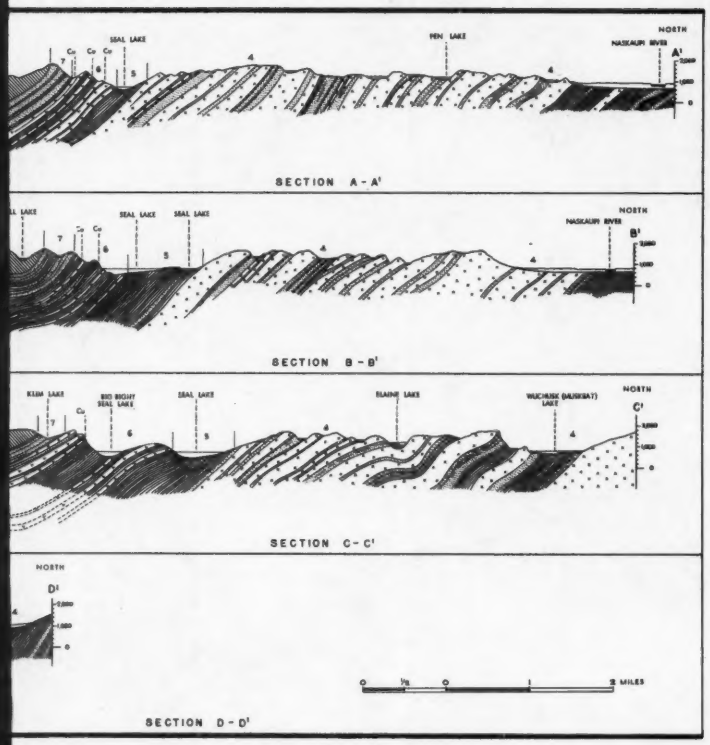


SECTIONS SHOWING GEOLOGY OF THE
(*See* Figure 3 and Plate 6 for

BRUMMER AND MANN, PLATE 5
Geological Society of America Bulletin, volume 72



Legend

PLEISTOCENE & RECENT

- Sand & glacial drift

SEAL LAKE GROUP (Proterozoic)

- UPPER RED QUARTZITE FORMATION – Red quartzites
- ADELINE ISLAND FORMATION – Red & gray shales and quartzites
- SALMON LAKE FORMATION – Basalt flows, diabase sills and red shales
- WHISKY LAKE FORMATION – Red shales and argillites
- WUCHUSK LAKE FORMATION – Diabase sills, quartzites, shales, chert, limestone
- BESSIE LAKE FORMATION – Chlorite–epidote schist, quartzites and minor conglomerates

BASEMENT (Archean)

- LETITIA LAKE GROUP – Tuffs, rhyolites, argillites; quartz-feldspar porphyry } intruded by syenite with amphibolite
- GRANITE – Hornblende & biotite granites with minor quartz–sericite and chlorite–epidote schists

TYPES

- Quartzites and conglomerates
- Shales, slates
- Limestone
- Diabase sills
- Basalt flows
- Schists
- Silicic volcanic rocks and fragmentals
- Porphyries
- Syenites
- Granites

MO = Mineral occurrence
Cu = Copper
Fault

CTIONS SHOWING GEOLOGY OF THE SEAL LAKE SYNCLINE, LABRADOR
(*See* Figure 3 and Plate 6 for location of sections.)

amounts of copper (0.03–0.05 per cent) and silver (0.01 ounces per ton). The close association of copper mineralization with the basalts and diabases suggests a genetic relationship.

Thin-section studies of the flows show small amounts of glass in a finely crystalline matrix. The basalt is composed of abundant fine laths of plagioclase (about 70 per cent) with a trachytoidal to subophitic texture. Coarser-grained varieties show the feldspar laths in a matrix of fine-grained augite with abundant secondary epidote, chlorite, and accessory iron minerals with leucoxene. Finer-grained varieties have a glassy matrix with abundant feldspar laths and iron ore. Much of the plagioclase has been completely saussuritized, and the pyroxene has been partly altered to chlorite and epidote. Many of the flows contain abundant white oval amygdules which are filled with calcite and minor quartz.

Most of the veins carrying copper mineralization are epidote with small amounts of quartz, chlorite, and calcite and minor chalcocite, bornite, chalcopyrite, pyrite, and molybdenite. Others contain abundant platy albite at right angles to the walls of the vein and hair-fine acicular actinolite. Bent crystals of chlorite and very fine-grained dusty inclusions are present in quartz-albitite vein material.

The basalts should be termed "metabasalts" where metamorphism and metasomatism have been active. Drill-hole intersections 400 feet below the surface indicate that saussuritization and general alteration of the basalts are not related to weathering.

"DIABASE": One of the best exposed of the medium-grained diabase and metadiabase sills caps many hills near Whisky and Salmon lakes.

The sills are coarser-grained and dark gray to green and lithologically resemble the basalt flows of the formation more closely than the coarse-grained Wuchusk Lake formation diabase sills. However, the diabases are generally coarser-grained and more massive than the basalts, lack the red or purple mottling and red oxidized contacts, and have chilled contacts. Pronounced saussuritization, epidotization, and chloritization of the material are common. Where they are prominent, the rocks should be termed metadiabase.

Most of the material contains only a trace of copper.

Adeline Island formation. The Adeline Island formation is separated from the underlying Salmon Lake formation on the basis of the absence of basaltic flows. The formation reaches a maximum thickness of about 1400 feet along the northern limb of the syncline (Pl. 5, section B-B′). Rocks of this formation are greenish-gray, black, and red to purple slates and shales, interbedded with pink and white quartzite and argillaceous quartzite.

Copper mineralization occurs in shear zones in the argillaceous rocks at the Adeline Island and Ellis prospects. The mineralization consists mainly of chalcocite and bornite with minor chalcopyrite and pyrite.

Upper Red Quartzite formation. The Upper Red Quartzite formation, the youngest formation of the Seal Lake succession, occurs along the axis of the Seal Lake syncline in a belt 14 miles long and 3 miles wide, from Ellis Lake in the west to north of the western end of Salmon Lake in the east. The base of the formation is placed at the first thick succession of clean, red quartzites. The writers estimate that the formation is at least 2600 feet thick (Pl. 5, section B-B′). As the top of the formation has been removed by erosion, the maximum thickness is unknown.

The formation is composed of massive to well-bedded red quartzites which show very little variation.

A series of anticlinal and synclinal folds (Pl. 4, fig. 1) in the area represent minor axes of drag folding along the major axis of the synclinorium. They are especially prominent in the Upper Red Quartzite formation. The axes trend east but develop a slight en échelon pattern from the southeast toward the west.

METAMORPHISM

The rocks of the Seal Lake group are for the most part fresh and are only slightly metamorphosed. The effects of metamorphism are strongest along the southern limb of the syncline where the rocks have been greatly compressed and deformed by the overthrust mass of older granite.

Within the southern mass of granite the degree of regional metamorphism is very slight, but the degree of dynamic metamorphism increases markedly toward the sheared contact between the overthrust granite and the Seal Lake formations. A zone showing a high degree of mylonitization, about 100 yards wide and traceable for miles, marks the contact. The mylonitized rock is composed of large augen of broken feldspar with grains of fractured quartz in a matrix of sericite and chlorite (Pl. 1, fig. 4). The rocks in this area present an excellent example of gradation from unaltered

granite to highly sheared mylonite over a width of about one-quarter of a mile.

Some of the remnant metasedimentary bands within the granite exhibit a higher degree of regional metamorphism (Pl. 1, fig. 1). Chlorite and epidote are abundant in most of the rocks and represent the greenschist facies. Hornblende and epidote are present in a few bands and are indicative of the higher amphibolite facies of metamorphism.

The arenaceous sediments of the area have all been altered to quartzites. Cross-bedding, ripple marks, and normal depositional bedding are preserved in these. Microscopic examination reveals a slight silicification of the individual grains in many of the quartzites. The bedding planes are generally well preserved and contain magnetite grains in abundance along the partings. The degree of metamorphism increases markedly along the crests of folds: this has resulted in partial or complete recrystallization of the quartzites. Many of the clastic structures have been partially obliterated in these areas of recrystallization and a simple mosaic structure developed. Some fine-grained quartzites in the Wuchusk Lake formation may represent recrystallized bands of chert.

The volcanic rocks of the Bessie Lake formation along the southern limb of the syncline have all been metamorphosed to chlorite-epidote schists of the greenschist facies and lack the amygdules, flow structures, and phenocrysts of feldspar which are still preserved in their northern counterpart, the Majoqua Lake formation.

The argillaceous rocks of the syncline also reveal a low grade of metamorphism which is represented by phyllite and slaty phyllite, with a few bands of slate. The cleavage of these phyllites and slates is oblique to the bedding in many places.

Most of the plagioclase of the Wuchusk Lake formation is highly saussuritized, although unaltered material is present in a few places. The saussuritization of the feldspar appears to be more complete along the southern limb of the syncline, with the formation of abundant epidote and chlorite of the greenschist facies.

The metamorphic effects of the intrusion and extrusion of mafic rocks have caused only slight metamorphic effects on the surrounding rocks. Some of the argillaceous beds directly in contact with the sills or volcanic rocks exhibit hornfelsic textures.

Some of the veins studied near the copper prospects show the effects of metasomatism. Albitite dikes with chlorite, actinolite, and scattered chalcocite, bornite, and chalcopyrite are common in the basaltic rocks of the Salmon Lake formation.

MINERALIZATION

General Statement

Most of the more than 250 showings of copper mineralization are restricted to an area of approximately 180 square miles between East Seal Lake and Adeline Lake, and from the northern shore of Seal Lake to Esmee Lake (Pl. 6). Most occurrences are associated with basaltic flows and diabasic sills of the Salmon Lake formation and of the Whisky Lake and Adeline Island formations.

A few niobium–thorium–rare-earth deposits and traces of zinc-lead-copper mineralization occur in the Ten Mile Lake area.

Association of Copper Mineralization with Rock Types

The copper showings range from a malachite stain to mineralized shears 20 feet wide and 200 feet in length. Basalt flows contain 36.5 per cent of the occurrences; diabase sills 15.5 per cent; contacts of basalt or diabase with adjacent metasedimentary rocks 25.4 per cent. The mineralization occurs either along the contacts or within a few feet of them; shales, slates, or quartzites 50 feet or more from the nearest mafic igneous rocks (basalt or diabase) contain 22.6 per cent of the occurrences.

Thus in 77.4 per cent of the occurrences the mineralization appears to be directly associated with basalts and diabases. Some of the remaining occurrences are several hundred feet above the nearest mafic igneous rocks.

Sulfides and Copper Minerals

The following minerals have been found either singly or in combinations: malachite, cuprite, native copper, chalcocite-native copper, chalcocite-bornite, bornite-chalcopyrite, chalcopyrite-pyrite, and traces of molybdenite, covellite, galena, sphalerite, and specularite. Chalcopyrite, chalcocite, bornite, and native copper are the most common, in that order.

Bornite, chalcopyrite, and pyrite are commonly associated with diabase sills, native copper, chalcocite, and bornite with basalt flows. All these minerals also occur in veins and shears in the adjacent metasedimentary rocks.

All mineral occurrences sampled returned

silver values, ranging from 0.25 ounce to 1–2 ounces of silver for each per cent of copper. Bornite-rich prospects seem to have higher silver values than those rich in the other copper minerals. Traces of native silver have been identified.

Some of the vein deposits have produced spectacular specimens of native copper, chalcocite, and bornite.

Gangue Minerals

The chief gangue minerals are: quartz, calcite, albite, epidote, and chlorite. Mineralization is commonly associated with zones of epidotization along fractures in diabase sills and basalt flows. However, not all epidotized fractures are mineralized.

TABLE 2. COPPER OCCURRENCES DESCRIBED FROM THE SEAL LAKE AREA, LABRADOR

Prospect	*Mineralization*
(1) Main showing	Native copper in red slates
(2) Adeline Island prospects	Chalcocite, bornite, and chalcopyrite in sheared schistose shales.
(3) Ellis prospect	Bornite and chalcocite in sheared shales
(4) North Whisky Lake prospects	Bornite and chalcopyrite in tension fractures in diabase sill Chalcocite in sheared diabase

Mode of Occurrence of Copper Mineralization

Copper mineralization commonly occurs along small shear zones, both in the igneous rocks and in the adjacent metasedimentary rocks. The widths of these shear zones are generally measured in inches; at Adeline Island prospect, however, the shear zones reach a width of 20 feet. Lengths up to 100 feet are common, and a few extend for a few hundred feet, but the majority are very short. The copper minerals occur with or without quartz gangue. The prospects at Adeline Island, Ellis showing, Main showing, and the No. 3 zone of North Whisky Lake occur along shear zones (See Table 2).

In addition, lenticular copper-bearing veins are found along tension fractures which commonly occur along the contact between the metasedimentary rocks and the intercalated mafic rocks. Few measure more than a few inches across. Many of the veins are short, but some can be traced sporadically for 100 feet or more. In a few places the adjacent wall rock is mineralized for a few inches on each side of the vein. The East Seal native copper occurrence and the No. 1 zone of North Whisky Lake are examples of this type of occurrence. Some of these deposits have produced specimens of native copper as much as 4 inches wide and 3 feet long.

The shearing was produced during folding by movement along the contact of beds of different competency, particularly between volcanic flows or diabase sills and the adjacent metasedimentary rocks. Tension fractures were produced in the more brittle beds along the lower contacts.

The main through-going faults and other crosscutting features do not appear to have controlled the mineralization.

Details of Some of the Copper Occurrences

The more promising copper occurrences were investigated by stripping, trenching, geophysical surveys, and diamond drilling where necessary.

MAIN SHOWING, SEAL LAKE (MINERAL OCCURRENCE 100): The presence of native copper first attracted attention to the area. Subsequent work uncovered many occurrences associated with the mafic igneous rocks of the area. The authors have examined one such deposit; the results are recorded hereafter.

The Main showing is on the south side of Seal Lake east of the narrow constriction which separates it from West Seal Lake.

Copper mineralization occurs in a band of red slates about 90 feet thick that is underlain by a red-green basalt flow and overlain by a diabase sill (Fig. 4). Native copper occurs sporadically along a zone 2–5 feet wide of parallel quartz-carbonate veinlets, each of which ranges in thickness from one-quarter of an inch to 8 inches. Most of the veins parallel strike and dip of the cleavage, but a few cut across the dip of the cleavage. Native copper along a single vein occurs in irregular masses up to 4 by 24 by 36 inches. Their distribution is erratic.

One locality on the lower contact of the overlying diabase sill has a number of quartz-carbonate veins which are erratically mineralized with small amounts of chalcocite, bornite, and chalcopyrite.

Native copper occurs at a number of other showings in the district.

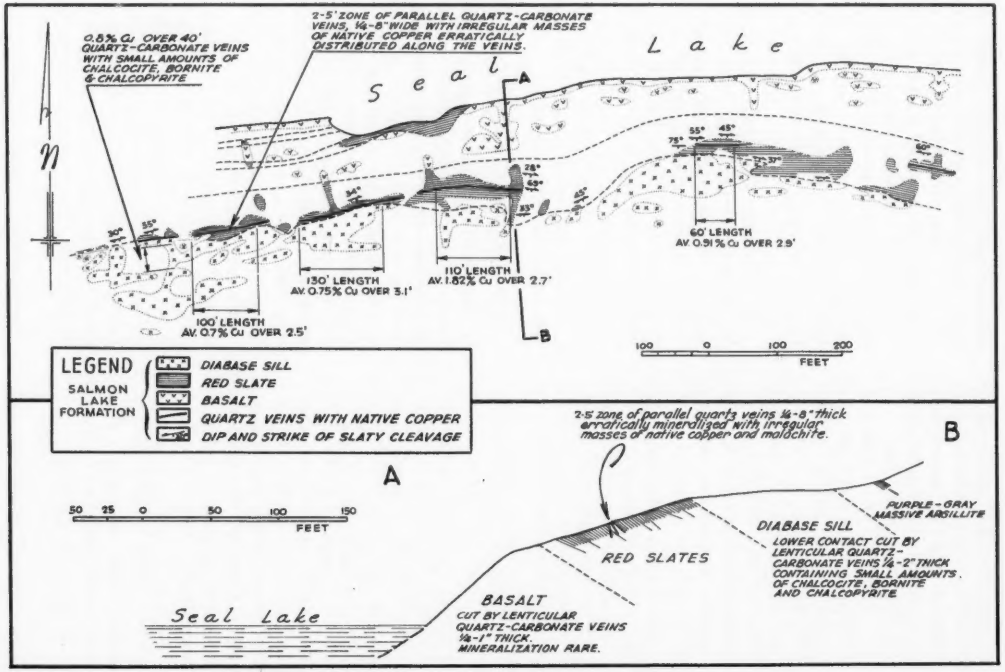
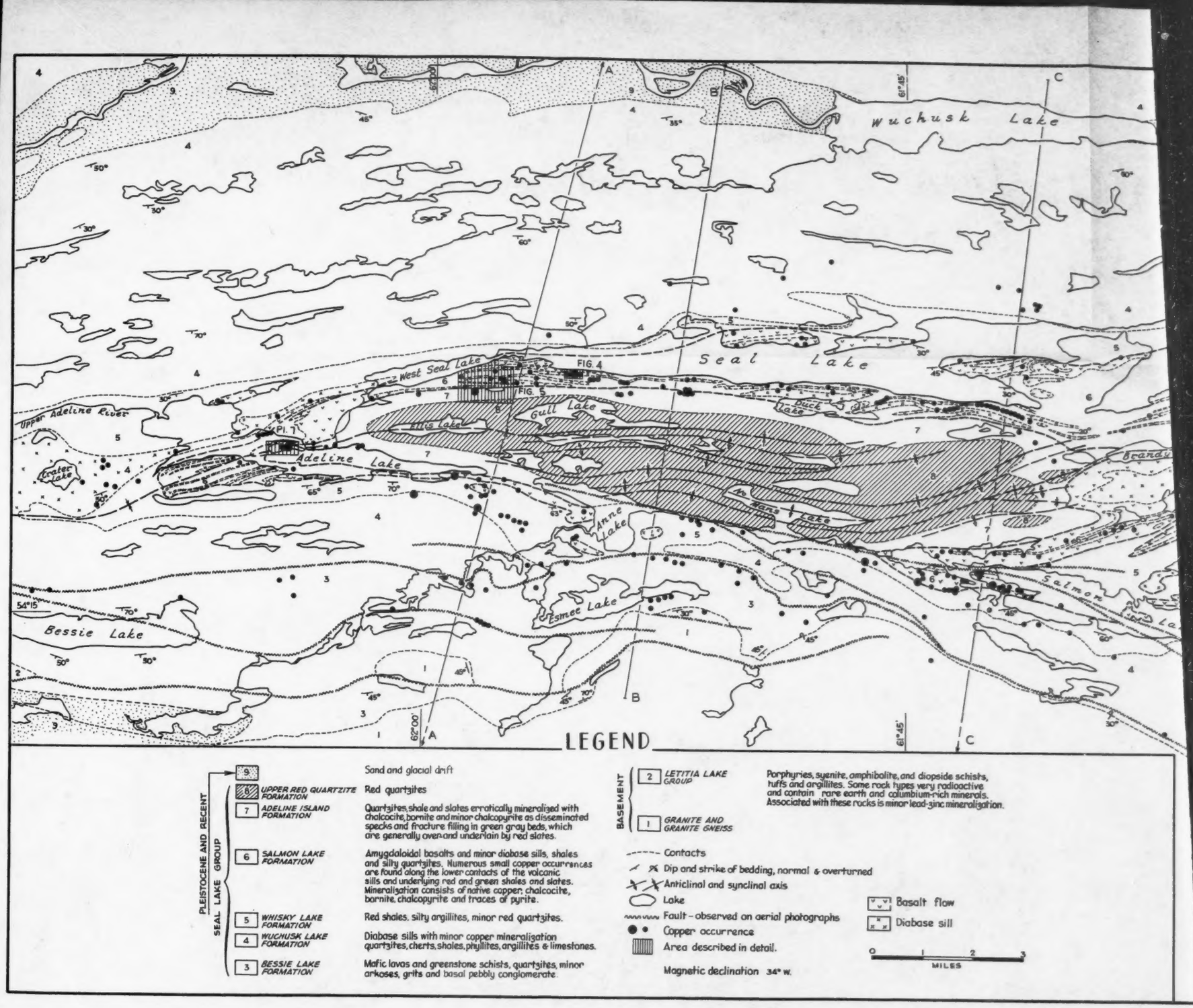


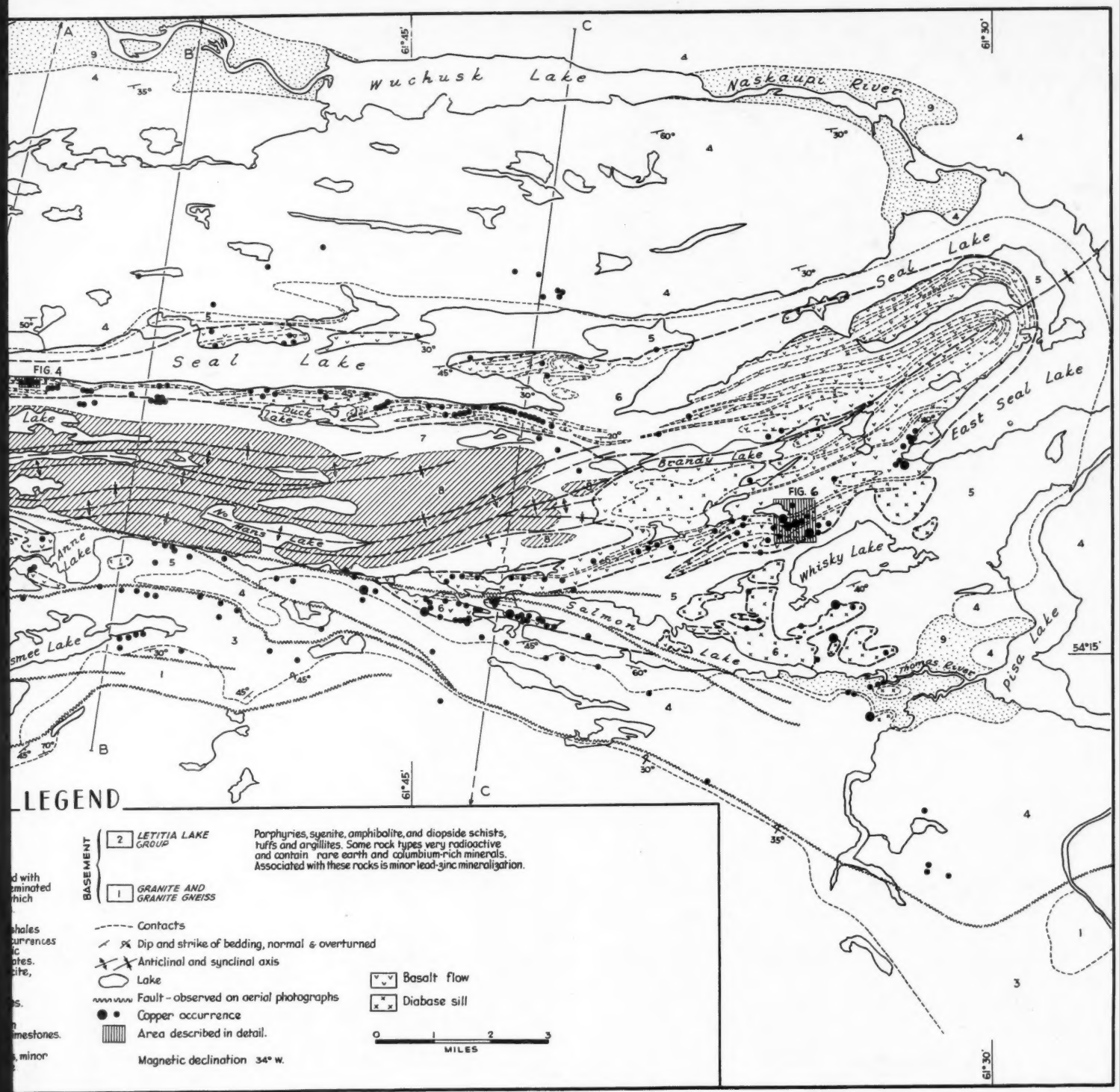
Figure 4. Plan and section of the Main showing. Section looking east


Wuchusk Lake
Seal Lake
West Seal Lake
FIG. 4
FIG. 5
Gull Lake
Duck Lake
Mills Lake
Upper Adeline River
Pl. 7
Adeline Lake
Crater Lake
Anne Lake
Esmee Lake
Bessie Lake
Salmon
62°00'
61°45'
A
B
C
LEGEND
PLEISTOCENE AND RECENT
SEAL LAKE GROUP
9 Sand and glacial drift
8 UPPER RED QUARTZITE FORMATION Red quartzites
7 ADELINE ISLAND FORMATION Quartzites, shale and slates erratically mineralised with chalcocite, bornite and minor chalcopyrite as disseminated specks and fracture filling in green gray beds, which are generally over-and underlain by red slates.
6 SALMON LAKE FORMATION Amygdaloidal basalts and minor diabase sills, shales and silty quartzites. Numerous small copper occurrences are found along the lower contacts of the volcanic sills and underlying red and green shales and slates. Mineralisation consists of native copper, chalcocite, bornite, chalcopyrite and traces of pyrite.
5 WHISKY LAKE FORMATION Red shales, silty argillites, minor red quartzites.
4 WUCHUSK LAKE FORMATION Diabase sills with minor copper mineralisation quartzites, cherts, shales, phyllites, argillites & limestones.
3 BESSIE LAKE FORMATION Mafic lavas and greenstone schists, quartzites, minor arkoses, grits and basal pebbly conglomerate.
BASEMENT
2 LETITIA LAKE GROUP Porphyries, syenite, amphibolite, and diopside schists, tuffs and argillites. Some rock types very radioactive and contain rare earth and columbium-rich minerals. Associated with these rocks is minor lead-zinc mineralisation.
1 GRANITE AND GRANITE GNEISS
Contacts
Dip and strike of bedding, normal & overturned
Anticlinal and synclinal axis
Lake
Fault – observed on aerial photographs
Copper occurrence
Area described in detail.
Magnetic declination 34° W.
Basalt flow
Diabase sill
0 1 2 3
MILES

PLAN SHOWING MINERAL OCCURRENCES AROUND SEAL LAKE, LABRADOR

Compiled by B. W. Hester and E. L. Mann. Formational contacts only are shown. Lava flows and sills are shown where related to mineral occurrences. Major structural featu

BRUMMER AND MANN, PLATE 6
Geological Society of America Bulletin, volume 72



HOWING MINERAL OCCURRENCES AROUND SEAL LAKE, LABRADOR

nal contacts only are shown. Lava flows and sills are shown where related to mineral occurrences. Major structural features are shown.

ADELINE ISLAND PROSPECT (MINERAL OCCURRENCE 132): This prospect is on an island in Adeline Lake.

Rocks on the island occur on the north limb at the western limit of the Seal Lake syncline. The south limb of the basin is under Adeline Lake. Two parallel flows of mafic volcanic rocks with interbedded red slates of the Salmon Lake formation which dip 50–60°S. (Pl. 7) underlie quartzites, green-gray schistose shales, and red and purple slates of the Adeline Island formation. The upper formation has been folded into tight, in places isoclinal, folds which, in general, are overturned with the axial planes dipping south. The troughs of the folds, in most places, are close to the surface and do not exceed 150 feet in depth. Considerable repetition of the beds results from folding. The tightness of the folding has produced considerable stretching and thickening of the beds, giving rise to boudinage structures in the quartzite layers with flowage in the adjacent slates. Variations in thickness of the beds together with interlensing of the sedimentary rocks make correlation difficult in places.

The structure consists of a number of separate folds of quartzites separated by slates. The northernmost folds are overturned to the south; the southern ones are more open. The quartzites exposed appear to be portions of the same bed which has been repeated by folding.

Mineralization is restricted to a green-gray sheared shale and schistose shale between greenish-gray quartzites and red slates and also to red to purple slates. Most of the sheared rocks occur along the slate-quartzite contact. The normal red or purple slate and the adjacent quartzites have been bleached to a green-gray color along this contact. The mineralized beds contain short lenticular, quartz-carbonate veins which are generally well mineralized. The veins extend into the adjacent red slates but are unmineralized there. The sulfides occur as irregular masses in the veins and as disseminations along the schistosity in schistose shales. Small fractures in adjacent gray quartzites are commonly mineralized with traces of sulfides. The minerals are chalcocite and bornite, lesser amounts of chalcopyrite, and traces of pyrite. One drill hole intersected sphalerite more than 1 foot in width. The color change from red to green gray in both slates and quartzites appears to be associated with the mineralization.

Surface sampling has shown the mineralization to be erratic, ranging from 5.1 per cent copper over 2.8 feet to 2.45 per cent for widths of 21 feet. The average along the best length is 2.1 per cent copper over 9.5 feet along a strike length of 250 feet. The eastern end of Adeline Island shows higher percentages of copper.

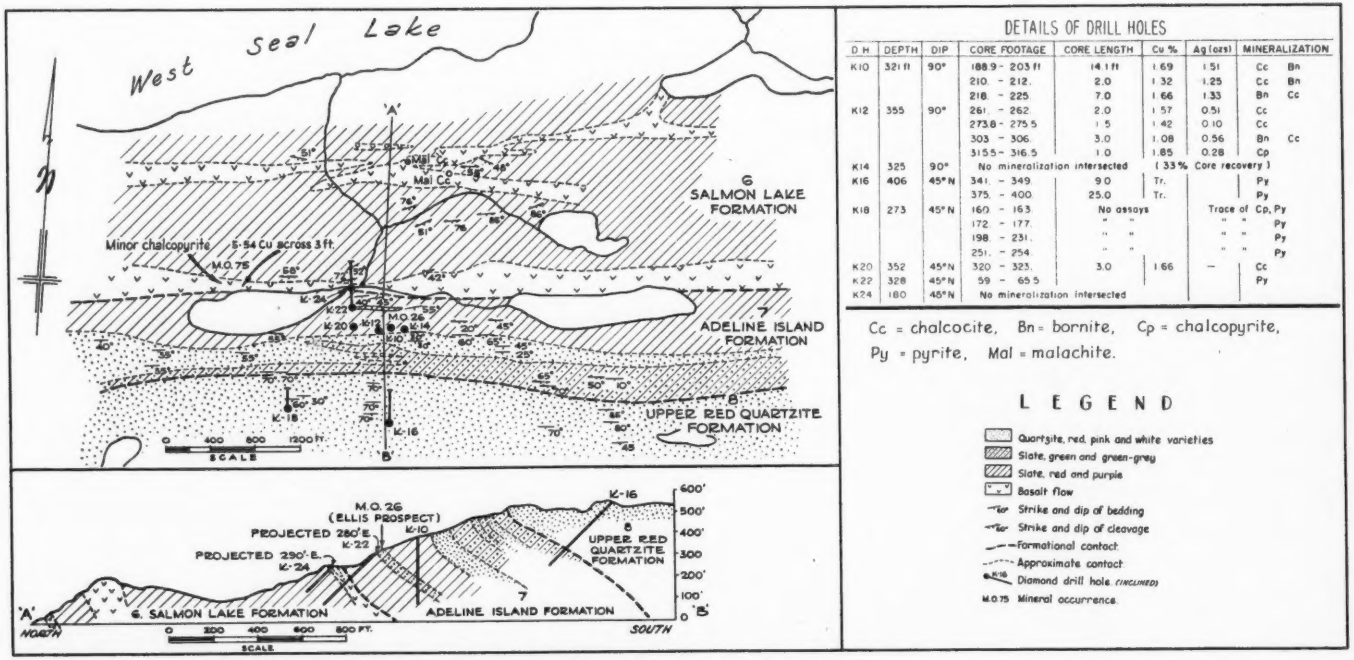
Diamond drilling has shown that only the roots of fold structures were preserved on the island. The mineralization is erratic and uneconomic, even though it is fairly widespread. The best value intersected was 4.70 per cent copper over a core length of 5.2 feet.

ELLIS PROSPECT (MINERAL OCCURRENCE 26): This occurrence is half a mile due south of the east end of West Seal Lake, between two small lakes on the side of a steep hill. Mineralization occurs along a shear zone in green shales in an intraformational shale-pebble conglomerate and in quartzites near the base of the Adeline Island formation (Fig. 5). The green shales occur interlensed with quartzite; both are overlain and underlain by red slates. The quartzites and green shales have been sheared, and boudins of quartzite occur surrounded by sheared shale. A series of closely spaced open shears immediately below the overlying red slates controlled the mineralization. This shear zone is lenticular and probably does not exceed 200 feet in length. Its width is variable with a maximum of 15 feet. Mineralization is mainly bornite and chalcocite, with a trace of silver. The highest values for copper are restricted to the shear zone; low values occur in the adjacent wall rock. Trenching has exposed a mineralized zone 177 feet long with an average width of 14.7 feet; the zone has 1.64 per cent copper and 2.40 ounces of silver per ton. Diamond drilling indicated the down-dip extensions of the mineralization to be irregular and discontinuous.

NORTH WHISKY LAKE PROSPECTS (MINERAL OCCURRENCES 64 AND 150): These prospects are north of Whisky Lake. The rocks in the area consist mainly of diabase sills which intrude red, green, and purple slates. The sills and metasedimentary rocks dip 20°–30°S. (Fig. 6).

Three mineralized areas have been located; of these, mineral occurrences 64 and 150 are the more important.

Mineral occurrence 64 (No. 1 zone) is along a dip slope of a diabase sill, immediately above the contact with green and red slates. The mineralization consists of bornite and chalcopyrite and minor pyrite and molybdenite occurring in small quartz-carbonate-albite veins along tension fractures in the sill. The diabase adjacent to the fractures is epidotized for a few



DETAILS OF DRILL HOLES

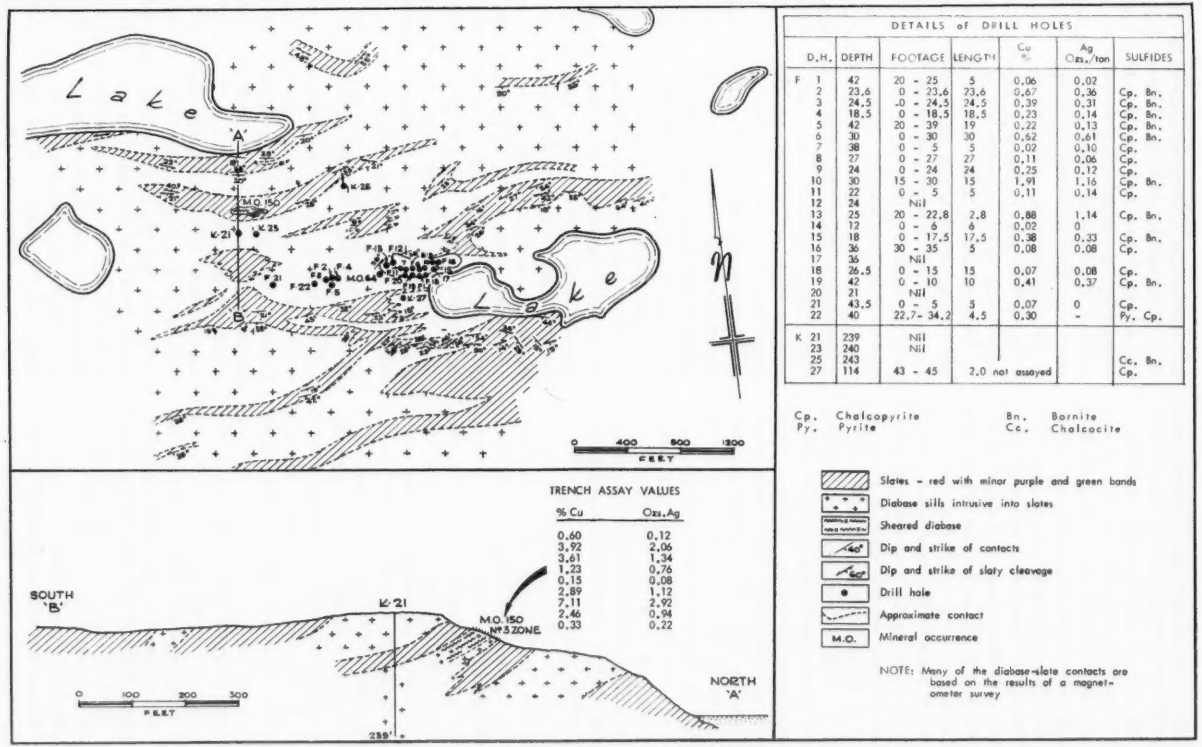
D.H.	DEPTH	DIP	CORE FOOTAGE	CORE LENGTH	Cu %	Ag (ozs)	MINERALIZATION
K10	321 ft	90°	188.9 - 203 ft	14.1 ft	1.69	1.51	Cc Bn
			210. - 212.	2.0	1.32	1.25	Cc Bn
			218. - 225.	7.0	1.66	1.33	Bn Cc
K12	355	90°	261. - 262.	2.0	1.57	0.51	Cc
			273.8 - 275.5	1.5	1.42	0.10	Cc
			303. - 306.	3.0	1.08	0.56	Bn Cc
			315.5 - 316.5	1.0	1.85	0.28	Cp
K14	325	90°	No mineralization intersected		(33% Core recovery)		
K16	406	45°N	341. - 349.	9.0	Tr.		Py
			375. - 400.	25.0	Tr.		Py
K18	273	45°N	160. - 163.	No assays		Trace of	Cp, Py
			172. - 177.	" "		" "	Py
			198. - 231.	" "		" "	Py
			251. - 254.	" "		" "	Py
K20	352	45°N	320 - 323.	3.0	1.66	—	Cc
K22	328	45°N	59 - 65.5				Py
K24	180	45°N	No mineralization intersected				

Cc = chalcocite, Bn = bornite, Cp = chalcopyrite,
Py = pyrite, Mal = malachite.

LEGEND

- Quartzite, red, pink and white varieties
- Slate, green and green-grey
- Slate, red and purple
- Basalt flow
- Strike and dip of bedding
- Strike and dip of cleavage
- Formational contact
- Approximate contact
- K-16 Diamond drill hole (INCLINED)
- M.O.75 Mineral occurrence.

Figure 5. Plan and section of the Ellis prospect



DETAILS of DRILL HOLES

D.H.	DEPTH	FOOTAGE	LENGTH	Cu %	Ag Ozs./ton	SULFIDES
F 1	42	20 - 25	5	0.06	0.02	
2	23.6	0 - 23.6	23.6	0.67	0.36	Cp. Bn.
3	24.5	0 - 24.5	24.5	0.39	0.31	Cp. Bn.
4	18.5	0 - 18.5	18.5	0.23	0.14	Cp. Bn.
5	42	20 - 39	19	0.22	0.13	Cp. Bn.
6	30	0 - 30	30	0.62	0.61	Cp. Bn.
7	38	0 - 5	5	0.02	0.10	Cp.
8	27	0 - 27	27	0.11	0.06	Cp.
9	24	0 - 24	24	0.25	0.12	Cp.
10	30	15 - 30	15	1.91	1.16	Cp. Bn.
11	22	0 - 5	5	0.11	0.14	Cp.
12	24	Nil				
13	25	20 - 22.8	2.8	0.88	1.14	Cp. Bn.
14	12	0 - 6	6	0.02	0	
15	18	0 - 17.5	17.5	0.38	0.33	Cp. Bn.
16	36	30 - 35	5	0.08	0.08	Cp.
17	36	Nil				
18	26.5	0 - 15	15	0.07	0.08	Cp.
19	42	0 - 10	10	0.41	0.37	Cp. Bn.
20	21	Nil				
21	43.5	0 - 5	5	0.07	0	Cp.
22	40	22.7 - 34.2	4.5	0.30	-	Py. Cp.
K 21	239	Nil				
23	240	Nil				
25	243					Cc. Bn.
27	114	43 - 45	2.0 not assayed			Cp.

Cp. Chalcopyrite Bn. Bornite
Py. Pyrite Cc. Chalcocite

Slates - red with minor purple and green bands
Diabase sills intrusive into slates
Sheared diabase
Dip and strike of contacts
Dip and strike of slaty cleavage
Drill hole
Approximate contact
M.O. Mineral occurrence

NOTE: Many of the diabase-slate contacts are based on the results of a magnetometer survey

Figure 6. Plan and section of the North Whisky Lake prospects

inches. The veins are up to 6 inches wide, averaging half an inch to 2 inches. The lengths are variable and do not exceed more than 3 feet. The veins are erratically distributed and are related to the proximity of the contact of diabase and slate. Occasional quartz veins are found in the underlying green slates, but these are unmineralized. Diamond drilling indicated erratic mineralization along a limited strike length, extending down to the underlying slate contact.

Mineral occurrence 150 (No. 3 zone) is 1000 feet northwest of mineral occurrence 64. Here a small local body of diabase occurs interfingered with red slates over a distance of 250 feet. Local shearing in the diabase has produced erratic and impersistent shear zones. Along these shears the diabase is mineralized with disseminated specks of chalcocite. Assays have returned values as high as 7.11 per cent copper and 2.92 ounces of silver per ton over widths of 5 feet. On the whole, mineralization is erratic, as is to be expected from the nature of the shearing.

Diamond drilling confirmed the surface observations that the mineralization is erratic and impersistent.

Conclusions

Traces of copper mineralization are widespread throughout the Seal Lake area, particularly in the upper formations of the syncline underlying the barren Upper Red Quartzite formation. Many of the showings have yielded spectacular specimens of native copper, chalcocite, and bornite. A considerable amount of time and money has been spent in seven years of intensive prospecting for economic copper deposits, but detailed investigation of these occurrences has indicated that the mineralization is erratic and uneconomic.

The copper-silver mineralization is mainly associated with the upper thin (100–200-foot thick) basalt flows and diabase sills of the Salmon Lake formation and the metasedimentary rocks of the overlying Adeline Island formation. Not all the diabase sills and basalt flows in the area are mineralized. The prominent Wuchusk Lake diabase sills and the extensive Bessie Lake amygdaloidal basalts, which occur lower in the sequence, are unmineralized, except for a few grains of pyrite and chalcopyrite in the diabase. It would appear that only the uppermost members of the igneous sequence of lava flows and sills contained sufficient copper and silver minerals to be concentrated during subsequent folding to give the small shear-filling and vein-type deposits found today. In some cases, the mineralization migrated along shears into the overlying Adeline Island metasedimentary formation and also, to a limited extent, into the adjacent underlying metasedimentary rocks.

Apparently no concentration of the heavy metals sufficient to produce economic ore deposits occurred. This may be due to the shallow depth of burial and low degree of metamorphism at the time of folding. In this case, the rock pile must have cracked open in numerous places during folding. If, before folding, burial had been deeper, the dispersed mineralization might have been funneled along major shears or faults to produce economic concentrations—other factors having been equal.

The occurrence of native copper and chalcocite in the basalt flows indicates a deficiency in sulfur. Cornwall (1951, p. 199) has suggested that the sulfur escaped as a gas during extrusion and crystallization of the lavas. The sulfur deficiency in the lavas contrasts with the situation in the diabase sills, where chalcocite, bornite, chalcopyrite, and minor pyrite are present. The sills retained their sulfur by virtue of their mode of emplacement.

The association of the copper mineralization only with the upper flows and sills in a thick sequence of mafic igneous and metasedimentary rocks suggests that the final outpourings of magma were enriched in heavy metals or came from a deeper sulfide-rich magma level.

Niobium–Thorium–Rare-Earth Deposits

Low grade deposits of niobium, thorium, and rare-earth minerals were discovered while the Letitia Lake area was being mapped. A reconnaissance geochemical survey had given anomalously heavy metal values in the Ten Mile Lake-Letitia Lake area (Brummer, 1960). General prospecting revealed small amounts of lead and zinc mineralization and a high radioactive background throughout most of the area. The Seal Lake group and the granites and anorthosites show low radioactivity. E. L. Mann discovered two radioactive zones, Mann No. 1 and No. 2, late in the summer of 1956.

MANN NO. 1 PROSPECT: This showing occurs within a large lenticular body of aegirine syenite near the gradational contact between the lower porphyritic rocks and the upper banded volcanic and pyroclastic rocks of the Letitia Lake group. The zone of high radio-

activity is concentrated along a banded complex of "gneissic" amphibolite and albite bands. The pronounced foliation of the rocks strikes about N. 80° E. and dips steeply south parallel to the general attitude of the Letitia Lake rocks. This banded complex represents a shear zone which contains large blocks of syenite in schistose amphibolite (Pl. 1, fig. 5). The results of a scintillometer survey delimited a zone of anomalous radioactivity (10× background) at least 1800 feet long by 500 feet wide, which closely follows the strike of the banded complex. The brilliant-colored oxidation products usually associated with uranium mineralization are absent here. It was not possible to sample the showing systematically, but samples from the zone of high radioactivity disclosed high values of niobium and small amounts of thorium, cerium, and rare-earth minerals. Assay values of grab samples of banded amphibolite show as much as 1 per cent Nb_2O_5; a zone of highly radioactive amphibolite showed a value of 2.30 per cent ThO_2. The rocks contain only trace amounts of uranium mineralization.

Small amounts of beryl and pink eudialyte have been identified from the zone. Extensive tests were run on specimens rich in niobium, but no recognizable niobium minerals have been identified, as the material is extremely fine-grained (R. J. P. Lyons, 1957, unpub. ms., Kennco Explorations, Canada, Limited).

MANN NO. 2 PROSPECT: A second radioactive zone is about 3 miles southwest of Mann No. 1. This zone strikes southwest with a steep southerly dip along the gradational contact between the porphyries and the banded volcanic and pyroclastic divisions of the Letitia Lake group. The zone consists of closely banded green pyroxenite, black amphibolite, and white bands of albite. The green bands contain abundant diopside or aegirine-diopside, whereas the amphibolite is composed of arfvedsonite. The rocks are of a skarn type, although no carbonate minerals have been identified in the vicinity. Aegirine syenite and fenitic syenite occur in the area. The banded skarn-type rocks and associated syenites can be traced sporadically for 20 miles to the southwest. Pink eudialyte is common in this portion of the area.

The radioactivity of the Mann No. 2 showing is less intense than that of the Mann No. 1 showing, and assay values give low-grade amounts of niobium and thorium.

GEOLOGICAL HISTORY

The geological record presented by the basement rocks of the Seal Lake area (granite, anorthosite, and the Letitia Lake group) is not very clear. Relict bands within the granite represent very old sedimentary rocks which have been engulfed. The Letitia Lake group represents a major period of volcanic activity. A thick succession of porphyritic flows and sills at the base is overlain by alternating bands of silicic lavas, tuffs, and pyroclastic rocks. The sequence has been intruded by aegirine syenite, but the age relationships of these intrusive rocks and the Seal Lake rocks are not known.

Deposition of the Seal Lake group was initiated on the peneplaned basement. The climate was probably relatively arid, judging by the fresh feldspathic matrix of the basal members of the Bessie Lake formation.

Prevalent cross-bedding and ripple marks in the Bessie Lake formation indicates deposition in shallow water. The well-sorted thick succession of clean quartzites has been washed under prolonged strand conditions.

Sporadic volcanic activity during deposition of the upper part of the Bessie Lake formation is represented by abundant amygdaloidal basaltic flows intercalated in the bands of quartzite. Subaerial extrusion is indicated by the absence of pillow structures and by oxidation of the flow tops. Vesicles and amygdules are abundant in most of the flows. Some of these volcanic rocks on the north limb of the syncline have been partially eroded. Disconformities are represented by bands of intraformational conglomerate which have been buried by later flows of basalt.

Shallow-water conditions still prevailed at the start of deposition of the Wuchusk Lake beds. Before long, however, the basin began to subside, resulting in deepening of the water. Finely banded cherts, shales, quartzites and algal limestone indicate that quieter conditions of sedimentation continued on until the close of the Wuchusk Lake deposition. Sea level fluctuated slightly, and periods of shallow-water deposition are represented by well-preserved ripple marks and aqueous cross-bedding in some quartzite bands. The algal limestone toward the base of the formation represents a similar period of shallow-water deposition.

The Whisky Lake formation was initiated with a period of terrestrial or shallow-water deposition represented by a thick succession of red shales and sandy argillites. Shallow-water deposition persisted until the onset of deposition of the Adeline Island formation.

The normal deposition of Salmon Lake shales

was interrupted by outpourings of basaltic lava flows, represented by at least four or five amygdaloidal basalt layers. The abundant small occurrences of copper mineralization in the area appear to be directly associated with these flows. Most occurrences are either within the igneous rocks or in the metasedimentary rocks in contact with them. The coarser-grained diabasic sills of the Whisky Lake and Salmon Lake formations and the diabase sills of the Wuchusk Lake formation appear to represent the intrusive phase of this volcanism.

With the onset of the deposition of the Adeline Island formation, the conditions of sedimentation changed from still-water conditions to shallow-water strand conditions which prevailed up through the deposition of the Upper Red Quartzite formation.

All the members from the upper part of the Whisky Lake formation to the youngest members of the Upper Red Quartzite formation are terrestrial in origin, as indicated by their pronounced red coloration. This indicates that the rate of subsidence of the geosyncline was about equal to that of deposition within it. The prominent bedding, ripple marks, and cross-bedding of the very clean, red quartzites point to deposition along the shore, with currents from the south, east, and west. Erosion has removed all record of further sedimentation within the Seal Lake sedimentary basin.

The deformation of the Seal Lake basin represents a major orogeny in the area. It resulted in the southern granite basement massif being thrust northward, compressing the rocks of the Seal Lake basin into a long narrow syncline with complete overfolding of the strata and prominent thrust faulting along the southern margin. This period of deformation is marked by regional and dynamic metamorphism.

There is no record of any post-Seal Lake sedimentary deposition in the area. The flat-lying sedimentary rocks of the Shipiskan plateau probably represent undeformed rocks of similar age to those of the Seal Lake syncline.

The rocks of the area are all of Precambrian age. The algal structures of the Wuchusk Lake formation suggest that the rocks of the Seal Lake syncline are late Precambrian (Proterozoic). No age determinations based on radioactivity are available for the rocks of the Seal Lake syncline. However, vein material containing lead and zinc mineralization in rocks of the Croteau Lake group from the Green Pond area, northeast of Seal Lake, has been dated by the lead-isotope method as 1685 ± 160 million years old (Cumming *et al.*, 1955, p. 66).

During deformation of the Seal Lake group, tension fractures developed in the more brittle basalts and diabases, shears in the adjacent shales. Copper minerals in mafic rocks of the Salmon Lake formation migrated into the tension fractures and shears.

Extensive erosion followed the deformation and probably persisted until the Tertiary, resulting in peneplanation of the area. Late Tertiary uplift initiated a period of rejuvenation in the river valleys. The area attained maturity before the onset of glaciation.

The area was extensively glaciated during the Pleistocene. A slight submergence of the continent during glacial times has been followed by an isostatic rebound of the margins of the continent. This period of emergence has been marked by active rejuvenation of the rivers in the area.

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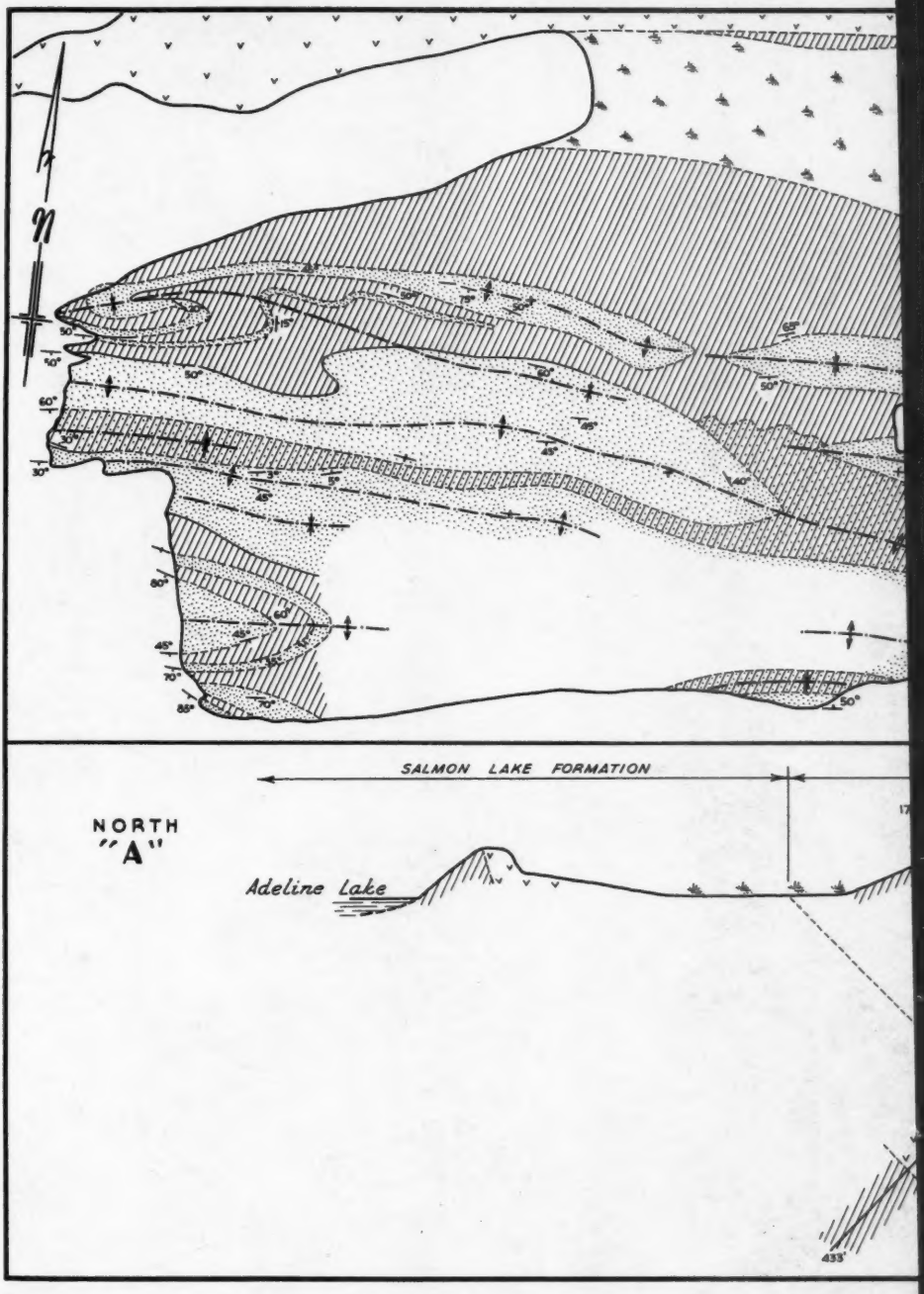
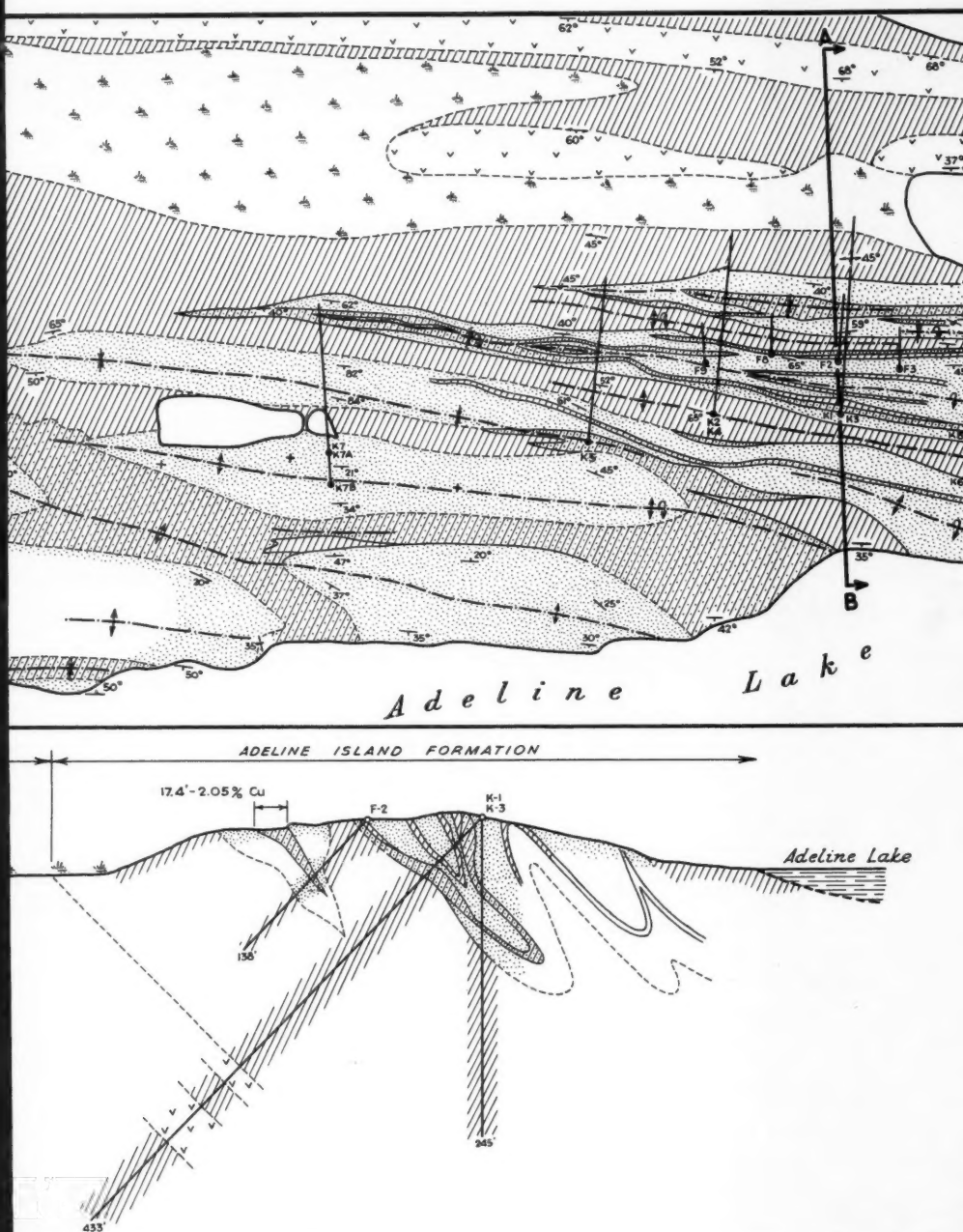
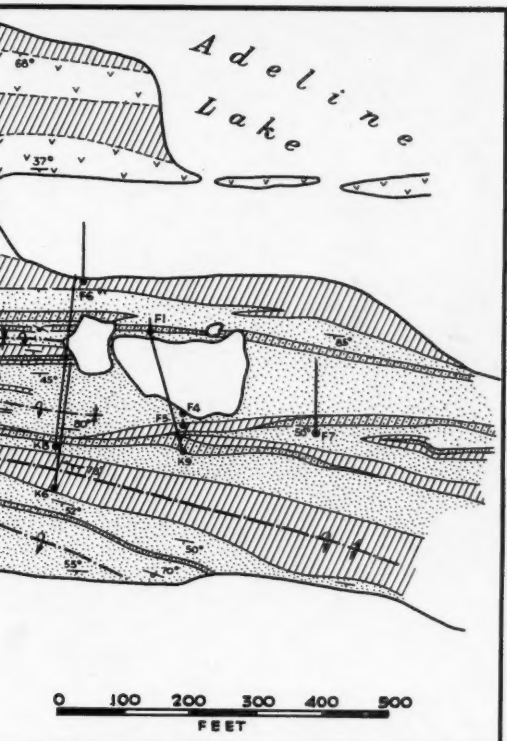


BRUMMER AND MANN, PLATE 7
Geological Society of America Bulletin, volume 72



PLAN AND SECTION OF THE ADELINE ISLAND PROSPECT, LABRADOR



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DETAILS OF DRILL HOLES

D.H.	Depth	Dip	Core footage	Core Length	Cu %	Ag ozs.	Mineralization
F1	50'	60°N.	20.0' - 23.2'	3.2' (44% Rec.)	0.99		Mal, Bn
F2	138'	45°N.	3.4' - 6.0'	2.6'	0.20		Bn
			61.2' - 63.5'	2.3'	3.70		Bn
F3	116'	45°N.	77.2' - 82.8	5.6'	2.0		Bn
F4	20'	45°N.	Abandoned				
F5	20'	58°N.					
F6	175'	55°N.	127.3' -136.0'	8.7'	-		Native Cu
F7	100'	45°N.	18.0' - 18.9'	0.9'	0.64		Mal, Bn
			83.5' - 86.0'	2.5'	0.24		Bn "
F8	90'	49°N.	2.2' - 3.9'	2.7'	0.70		Bn
F9	93'	45°N.	0 - 3.8'	3.8'	Tr		
			61.1' - 61.9'	0.8'	2.23		Bn
			61.9' - 64.6'	2.7'	Tr		
K1	433'	45°N.	9.0' - 17.0'	8.0'	2.45	0.32	Cc, Bn
			54.0' - 62.5'	8.5'	1.09	0.16	Bn, traces of Cp
K2	382'	43°N.	36.6' - 37.5'	0.9'	2.30	0.85	Cc, Bn
K3	245'	90°	48.5' - 60.0'	11.5'	1.85	0.24	Bn, Cc, Cp
			85.1' - 86.9'	1.8'	1.23	0.27	Bn, Cc
K4	421'	90°	139.2' -120.2'	1.0'	2.58	1.32	Cc
			139.2' - 141.3'	2.1'	1.45	0.66	Cc
K5	365'	45°N.	13.9' - 16.3'	2.4'	0.97	10	Cp
			48.5' - 49.5'	1.0'	No assays		Sphalerite
			137.6' -140.6'	3.0'	1.95	1.00	Bn, Cc
K6	275'	45°N.	61.0' - 64.2'	3.2'	1.32	0.03	Cc, Bn
			87.2' - 92.4'	5.2'	4.70	1.23	Cc, Bn
			125.5' -128.0'	2.5'	1.61	0.50	Bn, Cc
K7	27'	45°N.	Abandoned				
K7A	29'	53°N.					
K7B	373'	41°N.	191.3' - 198.7'	7.4'	0.01	0.01	Py
K8	336'	45°N.	36.0' - 37.0'	1.0'	0.97	0.25	Bn
K9	267'	45°N.	52.5' - 53.8'	2.0'	2.96	0.43	Cc
			112.0' - 114.0'	1.3'	0.13	0.10	Cc
	Total:		3974.0'				

Mal = Malachite
Cc = Chalcocite
Bn = Bornite
Cp = Chalcopyrite
Py = Pyrite

LEGEND

- Quartzites, red-brown and pink with minor green-gray varieties
- Sheared shale and schistose shale, green-gray, contains short lenticular quartz-carbonate veins, erratically mineralized with disseminated specks of bornite and minor chalcopyrite
- Slate, red and purple
- Basalt, red-green interbedded with slates
- Drill hole - inclined
- Dip and strike of metasedimentary
- Bedding horizontal, vertical
- Dip and strike of cleavage
- Anticlinal axis with overturned axial plane
- Swamp

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Erosion Surfaces and Emerged Beaches in Hong Kong

Abstract: In the Hong Kong area two high sea-level stages at 15 and 5–6 m P.D.[1] are clearly marked. These appear to be the result of eustatic movements. A similar well-marked bench at −11 m P.D. represents a Late Pleistocene low sea level. At least one other period of low sea level, in which base levels were more than 60 m below P.D., occurred. Benches at 22 m, 40 m, and 71 m are found as remnants around most of the coast and are probably marine. Summit leveling at 130 and 230 m is common and may be marine or subaerial. A valley basement now abandoned at 430 m represents an old drainage line lightly incised into a subaerial surface indicated by summit heights of about 460 m.

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INTRODUCTION

Topography and Structure

The Colony of Hong Kong is a dissected upland area, with predominant steep slopes, which has been drowned, probably by the Flandrian (*i.e.*, postglacial) transgression (Fig. 1). Infilling of the heads of drowned valleys is still at an early stage, and there is no general coastal lowland. Slopes of 20°–30° meet the sea in many sheltered areas, but where erosion is important they have been oversteepened into coastal cliffs.

The solid-rock history of the area ends in Mid-Tertiary with the deposition of the Red Beds, the typical basin deposits of South China. After this deposition, faulting, folding, and considerable thrusting occurred, and these earth movements may have laid the foundation of the present topography and dislocated any earlier erosion surfaces. Thrust movements were directed mainly to the south, and in several areas the Red Beds have been overthrust by earlier volcanic rocks and locally overturned (Ruxton, 1960).

Since this Mid-Tertiary movement the area appears to have been relatively stable, and there is no evidence of local earth movements in Pleistocene and Recent times. The Breccia Beds in Crooked Harbour, believed to be upper Pliocene, show no disturbance, nor do the Pleistocene and Recent alluvial and marine deposits. Thus in Late Tertiary, Pleistocene, and Recent times erosion, governed by various base levels, has affected the area.

Past Work

Kingsmill (1862) was the first to mention higher sea levels in Hong Kong; he referred to a light-colored semi-indurated deposit found

[1] Mean sea level is 1.1 m (3.76 feet) above local Principal Datum (P.D.) and at spring tide is 2.5 m (8.5 feet) P.D.

Geological Society of America Bulletin, v. 72, p. 1383–1394, 6 figs., September 1961

Figure 1. Place-name and topographic map of Hong Kong and the New Territories

up to 100 feet above sea level on Hong Kong island. He reported fossils in it, but unfortunately the exposures appear to have been removed or covered by buildings.

Heanley and Shellshear (1932) noted that the large bays of the colony are drowned intermontane basins; they observed shells and barnacles adhering to bedrock up to 3 feet above high water and assumed a slight relative retreat of sea level. For the nearby Pearl River delta Hubbard (1929) noted submergence of up to 200 feet but also evidence of considerable emergence.

Schofield (1940) recorded raised beaches and a slight historical recession of sea level. In 1942 he attempted a more widespread though speculative correlation and postulated large-scale tilting to explain the drowning of the area. Canton was the fulcrum around which the region from Tibet to the Phillippines had been tipped seaward. Dips in the quartz porphyry sheets which he cited are not universally regular and do not support the hypothesis.

Brock and Schofield (1926) postulated a submergence "five to six hundred feet greater than now," with "rock terraces particularly noticeable at about 480, 350, 200, 100 and 60 feet" (145, 108, 61, 30, and 19 m). Davis (1952), recording these views, also notes "wave cut platforms, terraces and old beaches at about 60, 40 and slightly under 30 feet" and suggests (p. 133) that from a position slightly lower than at present the land has undergone two complete oscillations and finally risen to the modern level. All these levels were given without supporting evidence and were not related to any stated datum. Heim (1929) estimated the subsidence to be about 200 m but also noted a 15-m terrace in the northeast of the colony. Chan (1948) studied isolated wave-cut platforms, 5–6 m above the high-tide mark, near Canton, but thought that the drowning was caused mainly by step faulting downward on the coastal side.

This brief survey of the literature suggests the need to present a review of the evidence and attempt a revised denudation chronology of the area.

The paper is based on field work around many parts of the coast line of Hong Kong and on map and air-photo studies and determination of heights by level and aneroid in the field.[2]

[2]The writer has given a detailed study of part of Lantau Island (1956, M.Sc. thesis, Univ. Bristol).

ACKNOWLEDGMENTS

Acknowledgments are gratefully made to the Research Grants Committee of the University of Hong Kong for financial assistance, and to B. P. Ruxton and Prof. R. W. Fairbridge for criticism and encouragement in relation to this work.

EROSION STAGES

Higher Erosion Surfaces

In the dissected region of Hong Kong there are no true higher erosion surfaces, but accordant summits provide the basis for tentative reconstructions. Tai Mo Shan, Tai Yue Shan, and Ng Tung Shan (Fig. 1), the three highest summits, reach very near the 1000-m level which is common in the plateau of the interior of South China (King, 1950).

Below this, many resistant hilltops on volcanic rocks terminate between 460 and 470 m, and there is a strong suggestion of summit leveling at this height (Fig. 2). On the eastern peninsula of Hong Kong both major peaks reach just to this level, and on north Lantau, south Lantau, and to the south of Tai Mo Shan; plateaus and hilltops are similarly accordant. A surface at this height could therefore have covered a large part of the area.

Apparently associated with it are the group of accordant heights on Hong Kong island and to the northeast. In the central part of the island five separate peaks reach 430 m and are flanked by peaks and plateaus at 470 m. A similar arrangement occurs to the northeast along a line which may have been a valley basement in the higher surface. This substantiates the hypothesis that the 460 surface is subaerial in origin. Possibly it is Late-Mid-Tertiary in age, and the stage was terminated by uplift followed by renewed downcutting which later began to develop the basins of the area. The 460-m summits are developed almost entirely on volcanic rocks, and the development of basins on sedimentary and granitic rocks postdates this stage (Berry and Ruxton, 1960).

Benches at 230 and 130 Meters

Erosion must have continued through a considerable period, because by the time the next clear stage in the evolution of the area is recorded the granite cupolas and sedimentary rocks had been pierced and at least some of the basins already formed. At about 230 m there is

Figure 1. Place-name and topographic map of Hong Kong and the New Territories

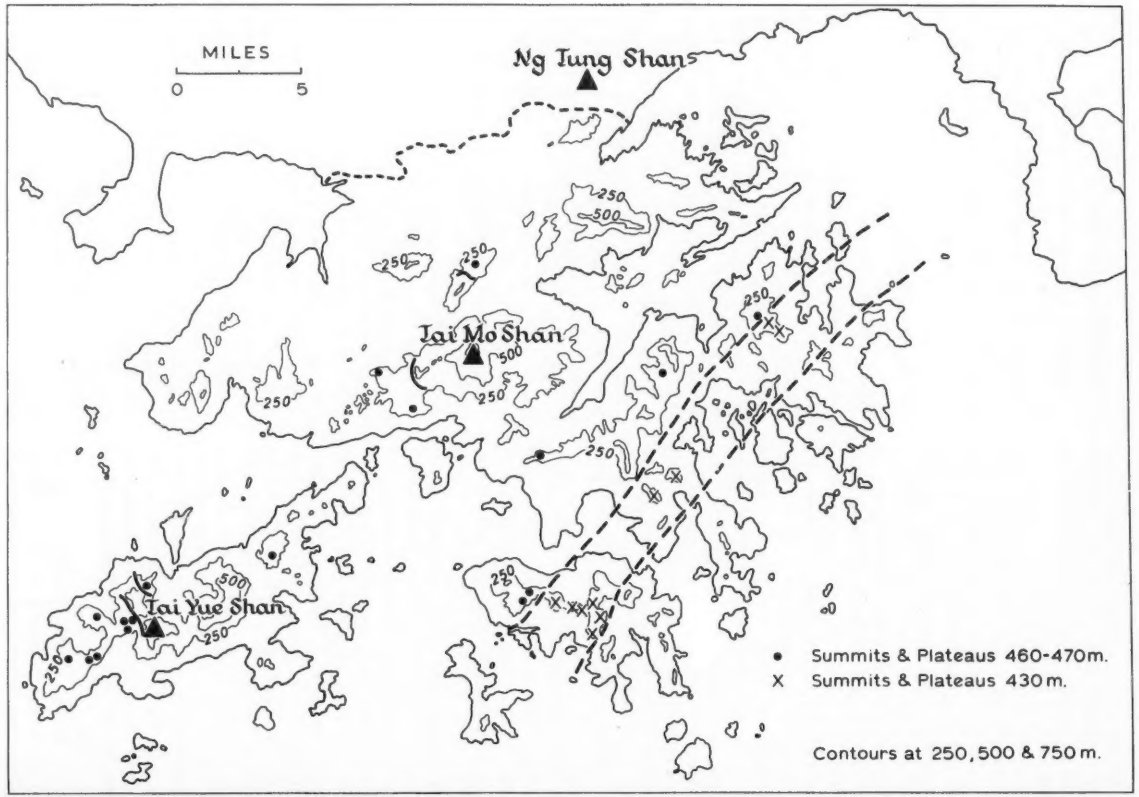

Figure 2. Remnants of higher erosion surfaces. The thick dotted line indicates the former valley basement at 430 m; the continuous thick lines show the junction between the 460-m plateau and higher ground

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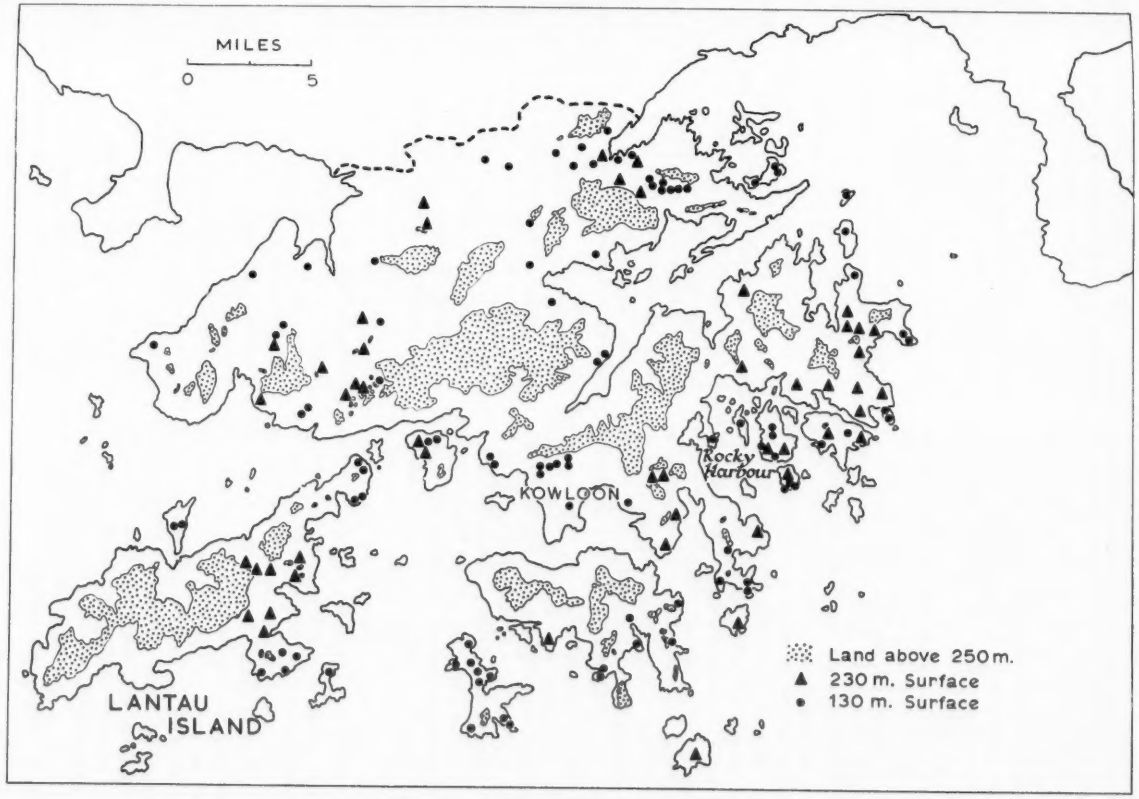

Figure 3. Distribution of remnants of the 230- and 130-m platforms

evidence of a further more distinct leveling, and remnants of the bench occur within the basin of the harbor area, in the Yuen Long depression, and on the slopes of the Mirs Bay depression. At this height flat-topped spurs and hilltops with gently rolling topography occur at many places (Fig. 3). Streams still have gentle gradients across the zone; in some places drainage is held up and marshes occur in the valley depressions which are graded to a base level of about 230 m P.D. The clearest development is around Rocky Harbour and on the Eastern peninsula, where the pattern of the bench can readily be reconstructed. Here and elsewhere it was at most a notching of the higher ground with sharp steep slopes on its upper margin. Weathering mantles on it are at least 15 m thick, though some erosion has occurred. No significant deposits have been found, and the origin may be subaerial or marine, though its greater development on the now exposed eastern areas points to the latter. From general considerations this is possibly Mid-Pliocene in age.

The stillstand was probably once more interrupted by wide-scale movements of land or sea, and the next episode is represented by spurs and hilltops leveled at about 120–130 m. Many islands terminate at this level; Lamma island has large areas with summits of this stage (Fig. 3), and the prominent peninsulas of the east coast are markedly beveled with rolling tops at this height. The widespread development of these benches indicates that the main outlines of the topography had now been developed, and several factors point to a marine origin. Benches with a rise to higher ground are marked by a steep upper slope which could be the old cliff line much degraded: the platforms are wider on the more exposed east, and they occur locally in such a pattern that origin by subaerial forces is unlikely. At its greatest development the bench was more than a mile wide; considering that it extends more than 500 m on hard rocks in the east, a long period of marine erosion must have occurred. It is probably Late Pliocene in age.

Lower Marine Benches

At 71 m, 40 m, and 22 m P.D. (232, 134, and 70 feet) distinct benches and dissected remnants occur around the coast line. Each group of platforms is terminated inland by a steep slope which leads in places to the next higher stage or to higher ground. At the seaward end most platforms are terminated by the coastal slope or cliff; their surfaces are smooth or lightly dissected showing remains of the former smooth surface. Most show a low seaward slope of less than 1°.

The 71-m bench is widespread in all parts of the coast; in places it remains only in the concordance of scattered hilltops, but in parts of north Lantau, the southern shore of Starling inlet, and on islands in Rocky Harbour, flat-topped spurs and headlands are continuous around wide stretches of the coast, and the form of the bench can be confidently reconstructed (Fig. 4). Across partly weathered granite rocks it may be up to 1000 m wide, but on unweathered volcanic rocks remnants in most places are 200–300 m wide. On granitic rocks considerable weathering has occurred since the formation of the bench, and the rock surface in many places is covered by 2–3 m of residual weathered debris. Preferential removal of this in some areas may have given rise to a number of benches at about 63–65 m P.D., although this may represent a distinct episode. On undissected remnants of this bench altitudes throughout the colony show no appreciable variation, from southern Cheung Chau to the north in Mirs Bay, and to the west in Deep Bay. In each of these and numerous other localities the levels have been checked by aneroid and tacheometric surveys and in some cases confirmed by government survey data. The 1 : 20,000 military maps of Hong Kong show clearly many remnants of the bench, and although these maps are not reliable in detail, altimetric data derived from them show it as a distinct landscape feature (Berry, 1956).

Confirmatory evidence of the base level at this height is found in some stream profiles. At Silver Mine Bay, Lantau, the main stream is graded in its upper reaches to the platform at 71 m which is represented by the hills around the bay (Fig. 4). As many of the streams in the colony are ungraded and small this confirmation is only locally available.

Most of the available data indicate a marine origin of the 71-m bench; its slope, terminations, and variable width in relation to exposure and rock type suggest that it was cut with a sea level about 71 m above P.D. and with topographic conditions much as they are at present. The extent and distinctiveness of the stage point to relatively long standstill, which cannot be dated on local evidence but is probably early Pleistocene.

After this period of stability and marine erosion changes of base level again occurred.

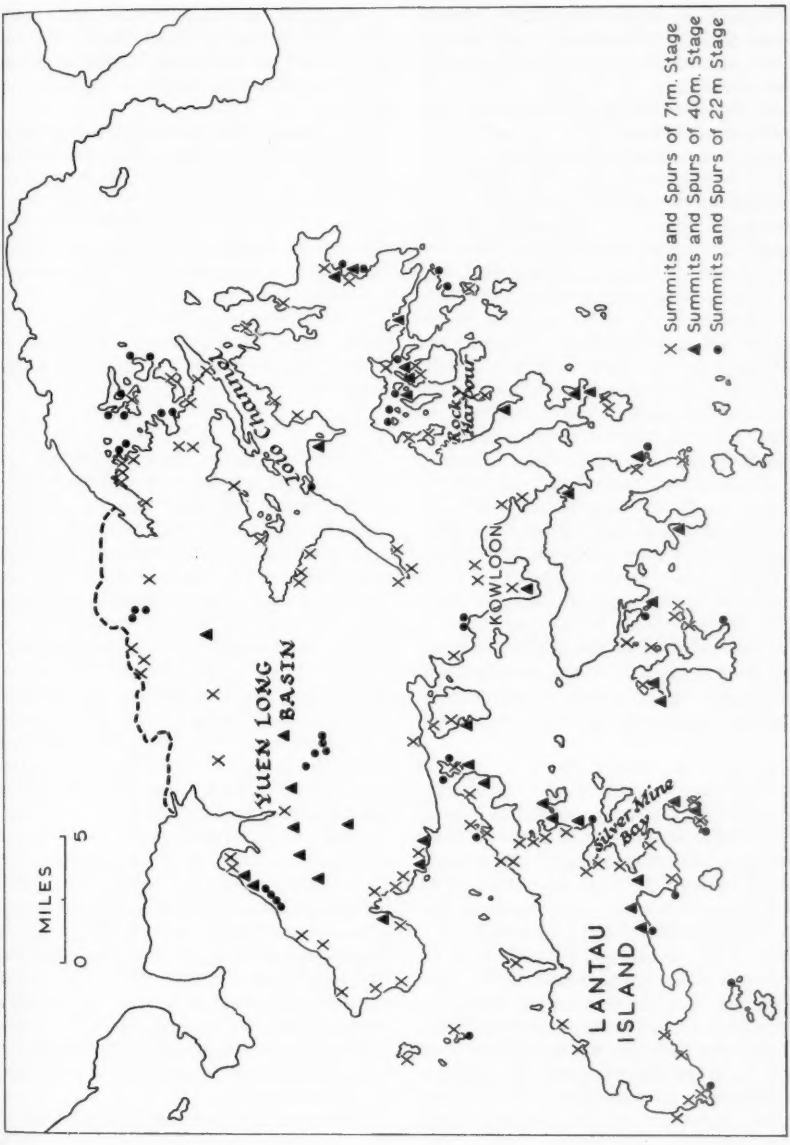

Figure 4. Distribution of remnants of the 71-m, 40-m, and 22-m platforms. Symbols offshore indicate remnant on nearby land

Whether just a simple downward relative movement of sea level took place, or whether the 40-m stage was the next peak after a series of lower oscillations, is not clear. The 40-m group of platforms are just as widespread but are usually smaller than the 71-m group occurring even in the inner bays and inlets of the coast line. Many are cut across hard bed rock but are well developed only in areas of weathered rock; this suggests a much shorter period of erosion than in the preceding stage. Because of their occurrence in scattered localities they have not been completely mapped, but wherever detailed studies have been made the bench is common. On Lantau (Berry, 1956) it occurs around much of the coast; it is particularly distinct and well developed across solid rock near Cheung Sha on the exposed southeast of the island, where its upper stretch is covered by slope-wash deposits. On Hong Kong island C. L. So (Unpublished data) found at least 12 distinct remnants on a coast generally unsuitable for excavation and preservation of benches of this type. Many more examples, few studied in detail, occur around the lowlands of Yuen Long area and in the southern shores of Mirs Bay. No beach deposits have been found in situ on this bench, and the evidence for its marine origin rests on altimetric data carefully collected from widely scattered areas and its distribution around much of the present coast line.

At 22 m P.D. another distinct bench is found, smaller but, because it is younger, better preserved in many areas. In solid rock it is at most 100–200 m wide, but in weathered rock it may be up to 400 m in extent. It is common around the colony but has been studied in detail only in Lantau, Hong Kong island, the Castle Peak-Kowloon coast line, and in some localities in the northeast. In each case its characteristics, height, and weathering features are similar. At Lai Chi Kok a particularly detailed study was made as the terrace was being removed for housing development. It had been cut across the weathered zones of a granitic rock, and since its formation some further weathering in relation to the new surface had occurred (Ruxton and Berry, 1957). The streams extended across the terrace after the retreat of sea level, and their new courses were in harmony with the greatest slope of the terrace surface, related to greatest fetch for locality. On Hong Kong island the bench is well developed on solid rock at Shekko where it is the only extensive locality for housing development on the otherwise steep eastern shore line. In the northeast many small islets show almost perfect leveling at this height, and a cave on Crescent Island was formed during this stage.

The evidence for these series of benches is consistent with the theory of a marine origin and suggests a relationship with the changing Pleistocene sea levels. The lowest bench may be linked with Monasterian benches found elsewhere in the Pacific, and post-terrace weathering suggests a long time interval between the first and last of these three stages.

In comparison with Pleistocene events elsewhere these stages probably represent the peaks of oscillations in level rather than halts in a steadily lowering sea.

Low Sea Levels

From evidence in Hong Kong only two stages of low sea level may be recognized, though others may have occurred. Drowned platforms are made more difficult to trace by the sedimentation over nearly the whole submarine area. The main outline of the present pattern of islands and peninsulas is obviously of subaerial origin, and the present bays and rias owe their origin to stream development at lower sea levels than at present. The depth of former base levels is difficult to determine. Borings at Yuen Long have reached more than 60 m below sea level before reaching bed rock, and in the deeper parts of Hong Kong harbor bores down to 35 m below sea level have not entered the old land surface.

Locally within Hong Kong waters greatest depths occur where tidal currents have prevented deposition. Soundings average 26–35 fathoms in narrow channels, and the greatest depth recorded, more than 43 fathoms, is between islands of the Pui Toi group, but some tidal scour is possible here. All lines of evidence suggest that stream activity at one period of low sea level was working at about −60 m P.D., although the base level may have been considerably lower than this if the shore line lay well to the east.

Numerous boreholes show the old land surface beneath marine deposits on the harbor floor. Much of it is weathered material resembling Ruxton and Berry's (1957) zone I or II of the weathered debris now found above sea level. Weathering depths of zone I more than 15 m thick have been found, but zone II has been little penetrated. The thickness of the upper zone indicates a considerable period of

subaerial weathering before the submergence.

Records of more than 900 bores in the harbor show that one third struck the old land surface between 10 and 12 m below Principal Datum (11-13 m below mean sea level). There is thus a marked buried platform at this level, related to a low sea level of some importance (Fig. 5).

Emerged Beaches and Associated Wave-Cut Platforms

Below the 22-m stage evidence of the high sea levels is found both in erosional and depositional forms. Emerged beaches occur at about 15 and 5–6 m P.D., but as Principal Datum is about 2.7 m below high-tide mark the relative changes of level are 12–13 m and 2.5–3.5 m (40 feet and 8–11 feet).

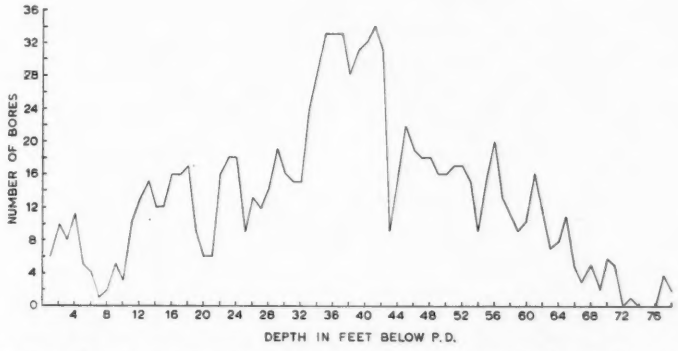

Figure 5. Graph to show altitudinal data of the old land surface below sea level. The marked peak between 33 and 42 feet represents a drowned platform.

The most distinctive emerged beaches of the upper group fringe much of the coast between Black Point and the north of Mong Cheng peninsula (Fig. 6). They are now being eroded by the sea at many places, showing a 10- to 20-foot section of fine-grained sand and silt, but the relatively sheltered nature of this area has resulted in the preservation of the beach here. Actual beach deposits are rare, as they are covered by slope wash at the lower part of the old cliff, clearly visible though now degraded. On exposed coasts remains of beaches of this stage are not common, but rock platforms and breaks in profile of the coastal slope at this height occur in many places. In the northeast, Robinson Island and Sai Ap Chau were leveled at this stage, and the "three top hats" in Three Fathom Cove are also about 15 m P.D. Many other probable examples have not yet been confirmed by field mapping. At the time of this sea level much of the low-lying land of the colony was submerged, and the sedimentation of these areas dates from this stage. The probable coast line indicated in Figure 6 is partly based on altitudinal data but is confirmed by marine fossils found in the Sheung Shui basin (Davis, 1952) and by boreholes through sediment at Yuen Long.

Soil studies in Laffans plain indicate that at this period drainage was poor, and swamps now typical of the low-lying coastal areas were found. Although dissected by streams and partly covered by alluvial deposits the basins of the northwest were infilled during and after this sea level. Around their margins distinct terraces exist on the lower hill slopes and on the streams which have been rejuvenated by the retreat of the sea from this height.

The 5- to 6-meter emerged beach deposits are found in many areas, although a lot of evidence has been destroyed by sand removal for construction work. On many granite beaches a double beach is found in the form of two sand spits built across a drowned valley, one at the beach and the other well inland.

The inner beach now surrounded by river and lagoon alluvium is 2–4 m above the present beach and in many places consists of iron-stained partly cemented sand, much of which is better rounded and graded than that of the modern beach. Where old and modern beaches occur together without a prograded stretch a similar sand and height distinction can be made.

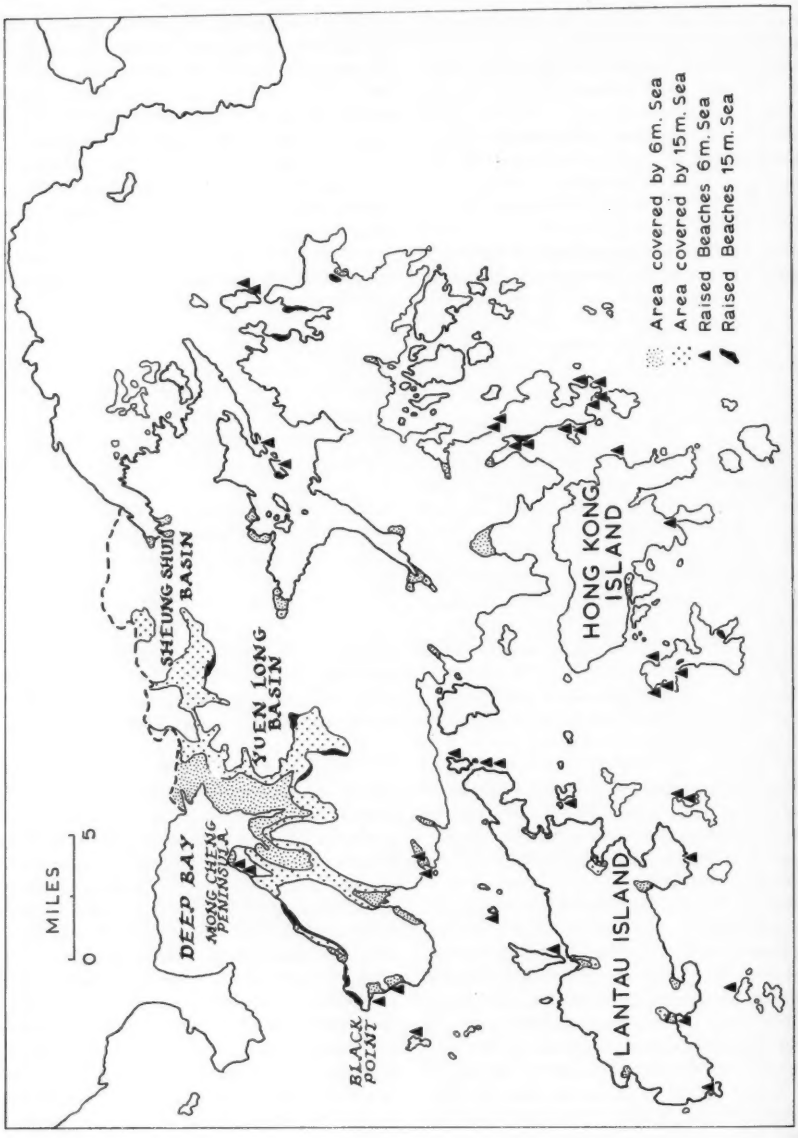

Figure 6. Former shore lines and emerged beaches at 15-m and 6-m stages

Schofield (1942) noted that only protohistoric remains are found in the higher beaches. He also distinguished a second historic slightly raised beach, with archaeological remains, which lies behind the modern beach. However, the tidal range and storm waves of the region make this difficult to prove.

Several tombolos in the area were partly formed at this higher level; the spit at Cheung Chau and the tombolos at Stanley and elsewhere were initiated at this time (Berry and Ruxton, 1957). In the lowlands of Deep Bay the approximate shore line of this period can be reconstructed (Fig. 6).

The streams in the lowlands and small bays are incised 1–3 m into their beds as a result of the fall in sea level from this stage. Sedimentation in the Mai Po and Yuen Long area was probably much the same as at present; fine silts and muds form all but the upper shore sediment, and the mangrove swamps would greatly help such deposition.

Water layer weathering and stripping of the basal surface make it difficult to distinguish benches formed during this stage from those formed at the present level (Johnson, 1933; Bartrum, 1926; Ruxton and Berry, 1960). However, benches related to the higher sea level are found; Heanley's mosquito notch, a bench noted on Cheung Chau (Personal communication), and the platform around Tolo Channel are probably of this age, although in the latter case tidal waves after large typhoons occur in this narrow channel.

A cave on eastern Hong Kong island provides clearer evidence. It is more than 15 m high at the head of a long narrow inlet and has two sections, a lower cave about 2.5 m high and an upper cave 11 m high separated by a ridge of rock. The lower section is now being excavated by storm-wave erosion at the base and by collapse of the roof; the upper section was formed similarly when the sea level was about 3 m higher than at present. This higher section is still partly filled with shingle and boulders cemented with calcite.

Figure 6. Former shore lines and emerged beaches at 15-m and 6-m stages

DISCUSSION

Correlation of these conclusions with other parts of the East Asian coast is difficult; few studies have been made, and many areas have been affected by tectonic movements. In Hong Kong the recognition by Heim (1929) of the 15-m terrace is confirmed, and the observations of Heanley and Shellshear (1932) and Schofield (1940; 1942) with regard to the raised beaches are generally supported, but the levels given by various authors for the higher stages must be discarded. Raised beaches are probably found over much of the southeast Chinese coast in spite of the statement to the contrary by Charlesworth (1957). Lin Kwan Te (1938) records several observations of raised marine phenomena of varying elevations. Caves related to a similar stage to the Hong Kong 5-6-m level were recorded near Foochow, and terraces up to "14 meters above the sea" were also found. Maglioni (1938) also found archaeological implements in raised beaches similar to those of the 5- to 6-m stage in Hong Kong, and Andersson (*in* Schofield, 1940) found similar forms in Indochina. Thus, whatever earth movements may have affected this coastal area previously, stability of land and movement of sea level seems dominant in recent time.

Dating of material from the raised beach in Hanoi (Unpublished data), by carbon-14 dates on contained shells, suggests an age of 3125 years ± 150 as the period of the beach formation, and this may be generally correlated with the postglacial terraces elsewhere (Fairbridge, 1950; personal communication). Correlations for the terraces above this height must be very tentative, owing to the great distance from other centers of work. A 45-foot (15-m) terrace has been recorded by Stearns (1945), but his intermediate 27-foot terrace has not been recognized in Hong Kong. Stearns (1945) does not mention a 22-m terrace, but this is included by Tester (1950) and may be Monasterian in age. For the higher sea levels no widespread correlations can be made until much more work is done in the southeast Asian area.

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Stratigraphy of the Ashford Area, Southern Cascades, Washington

Abstract: Upper Eocene to Oligocene(?) rocks of the Ashford area, southern Cascades, Washington, include terrestrial sedimentary types derived from plutonic and metamorphic sources and dominantly andesitic volcanic flows and volcaniclastics. The sedimentary rocks crop out along the Nisqually River near Mount Rainier National Park and intertongue with andesitic volcanic rocks. At least 10,000 feet of andesite, basalt, and dacite, chiefly in the form of volcanic breccia, conglomerate, sandstone, and shale, with some thick flow units, overlies the sedimentary rocks of the Nisqually River. This series is lithologically divisible into lower, middle, and upper parts.

The sedimentary rocks are lithologically similar to those in the Puget group, and the volcanic rocks to those in the Keechelus andesitic series, but correlations are uncertain. New formational names are not warranted without more detailed studies

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INTRODUCTION AND ACKNOWLEDGMENTS

The Ashford area, enclosed approximately by long. 121°40′ W. to 122°00′ W. and lat. 46°32′ N. to 46°47′ N., lies within the heavily vegetated western foothills of the Cascade Mountains between the Cascade crest and the Puget Sound lowland (Fig. 1). Mount Rainier National Park forms the northeastern border.

Volcanic rocks of the area are part of an extensive Tertiary volcanic complex which extends from north of Snoqualmie Pass into Oregon. Sedimentary rocks of the area lie near the southern end of a coal-bearing series which crops out extensively along the eastern side of the Puget Sound lowland. Correlations within the region are difficult because of abrupt lateral and vertical changes of the units, rarity of fossils, lack of traceable marker beds, and dense vegetation.

The writer mapped the area in the summer of 1953 and part of 1954; this paper represents part of a Ph.D. thesis presented to the University of Washington. Drs. Peter Misch, Howard A. Coombs, V. Standish Mallory, and Harry E. Wheeler of the University of Washington spent many hours in helpful discussion and criticism with the writer. Dr. Aaron C. Waters of Johns Hopkins University gave appreciated advice on the present paper. The paleobotanical determinations by Roland W. Brown of the U. S. Geological Survey have been extremely helpful.

STRATIGRAPHY

General Statement

Volcanic rocks (flows and volcaniclastics) of the area are dominantly andesitic. The sedimentary rocks (sandstone and shale) are composed of plutonic and metamorphic materials.

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All the rocks are terrestrial, and fossil leaves date them as late(?) Eocene to Oligocene(?). Dikes and sills, including diabase, diorite, basalt, and andesite, intrude the sequence. The larger intrusions tend to be dioritic, the smaller ones basaltic or gabbroic.

plosion. Epiclastic volcanic sandstone is composed of sand-size volcanic fragments which originate by epiclastic processes; volcanic conglomerate is composed of abundant, large (>2 mm), chiefly rounded, water-worn fragments; volcaniclastic refers to rocks with a predom-

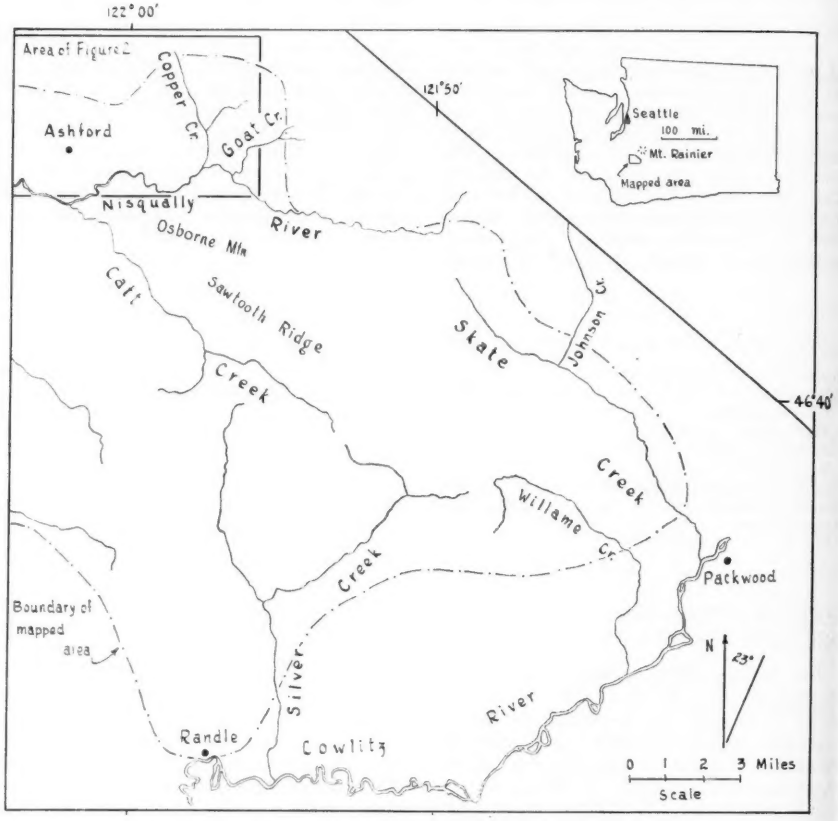


Figure 1. Index map showing location of Ashford area, Washington

Terminology

The writer uses "pyroclastic" as Wentworth and Williams (1932, p. 25) defined it, *i.e.*, as an adjective applied to rocks produced by aerial ejection of material from a volcanic vent. Volcanic breccia is composed of angular volcanic fragments larger than 2 mm, regardless of origin. Laharic breccia is a volcanic breccia resulting from mudflow, and pyroclastic breccia is a volcanic breccia produced by volcanic explosion. Epiclastic volcanic sandstone is composed of sand-size volcanic fragments which originate by epiclastic processes; volcanic conglomerate is composed of abundant, large (>2 mm), chiefly rounded, water-worn fragments; volcaniclastic refers to rocks with a predominance of volcanic fragments regardless of whether the rocks are pyroclastic or epiclastic (Fisher, 1961).

Sedimentary Rocks

Terrestrial sedimentary rocks composed of minerals derived from plutonic and metamorphic rocks crop out along the Nisqually River valley in the northern part of the mapped area (Fig. 2). They intertongue with volcaniclastic rocks. This sequence forms the lowest strati-

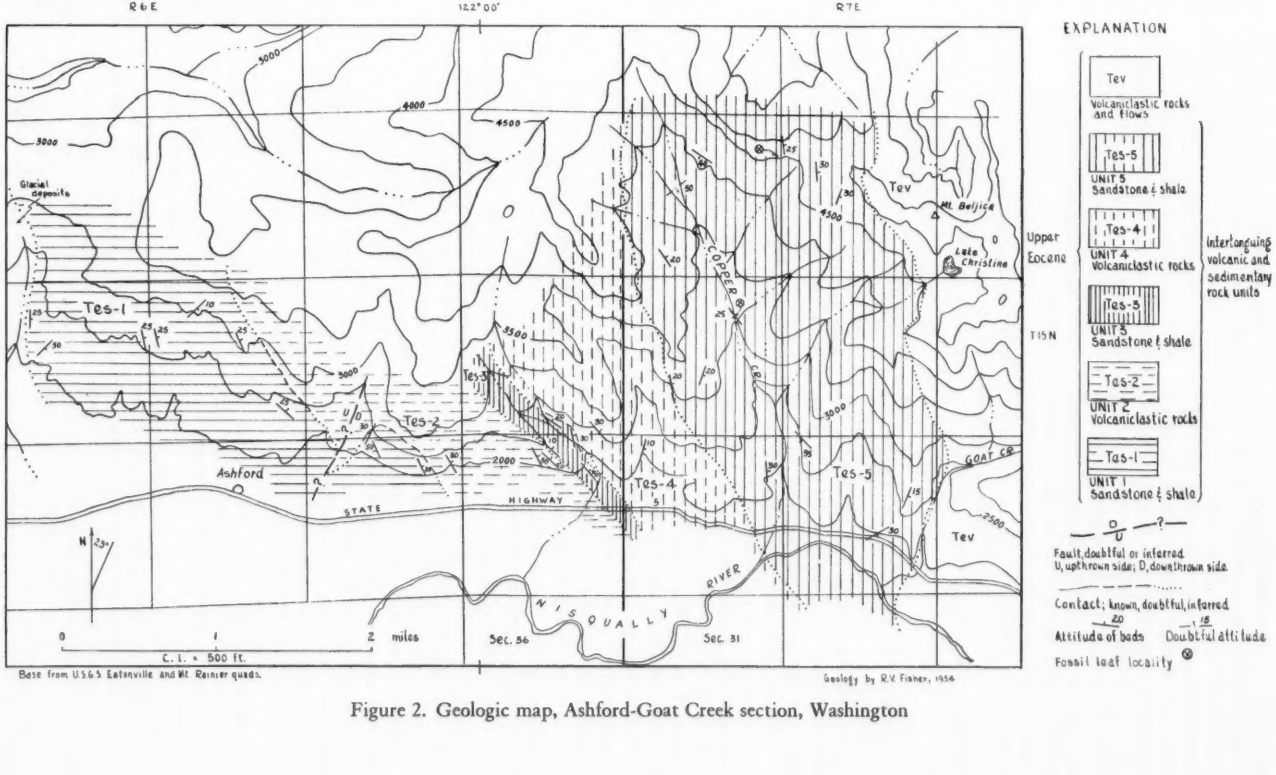


Figure 2. Geologic map, Ashford-Goat Creek section, Washington

graphic series of the Ashford area and is overlain by a thick series of andesitic volcanic and volcaniclastic rocks. Andesite flows and andesitic pyroclastic rocks of unknown stratigraphic position are contiguous with the sedimentary rocks on the west, but their description must await more detailed work.

Intertonguing of the sedimentary and volcanic rocks is best displayed along the north side of the Nisqually River from near Ashford to Goat Creek (Fig. 2). Here the following units can be recognized:

	Thickness (in feet)
(5) Sedimentary (youngest)	1900
(4) Volcanic	1500
(3) Sedimentary	400
(2) Volcanic	1800
(1) Sedimentary	3150+

The section totals more than 8750 feet. The following discussions of the units include petrographic descriptions of typical rocks.

UNIT 1: The rocks of unit 1 are friable to well indurated, massive to well bedded, medium- to fine-grained, light and dark gray to brown, muscovite-bearing arkose and sandstone with interbedded shale. Many of the sandstones show cross bedding, and some are ripple-marked. Plant fragments are locally abundant. One outcrop of sandstone contains rounded shale fragments. Interbedded silty shales, from 1 inch to several feet thick, are friable to fairly well indurated, fissile to massive and conchoidally fractured, medium gray to black and brown, and commonly contain muscovite. Some shales contain fossil leaves, and one shale outcrop contained poorly preserved brackish(?)-water pelecypods. Thin seams of lignite to bituminous coal occur within the sequence.

Near the upper contact, the sandstone becomes dark gray to black. The change upward from sedimentary rocks to coarse volcanic sandstones and conglomerates of unit 2 is rapid with little or no mixing of rock types.

Specimen 4: medium- to dark-gray, well-sorted, fine-grained sandstone from SE¼ sec. 22, T. 15 N., R. 6 E. about half a mile northeast of Ashford near upper contact of unit 1.

Average grain size 0.10 mm. About 35 per cent quartz (some strained), 5 per cent plagioclase (An_{26} to An_{29}), 1 per cent potassium feldspar, 10 per cent biotite (some altered to chlorite, many grains bent around resistant grains), 10 per cent muscovite (many grains bent), 5 per cent chlorite (some appears primary, most is altered, mainly from biotite), 2 per cent chert fragments, <1 per cent opaque minerals (magnetite?), 30 per cent matrix (dirty-brown, clay-size material and secondary calcite)

UNIT 2: Unit 2 is composed of well-indurated, massive, medium-gray, gray-green, green-brown, and brown volcanic breccias and sandstones. Coarse-grained rocks near the base grade upward to fine-grained rocks. The lower contact can be seen in two places from State Highway 5: on the steep valley slope north of Ashford, and on the steep valley slope about 1 mile east of Ashford in S½ sec. 26, T. 15 N., R. 6 E.

The volcanic breccias contain angular to well-rounded cobbles up to 4 inches, most of which are angular to subangular, black, gray, green and/or red andesite porphyries, none of which are vesicular. The matrix is volcanic sandstone with a few disseminated carbonaceous fragments.

The volcanic sandstones are medium- to fine-grained and in places are difficult to distinguish from flows, but fragmentary feldspar, occasional andesite pebbles and faint stratification planes distinguish them. They are dark green to medium gray green and usually very well indurated.

An andesite flow was found in SE¼ SW¼ sec. 23, T. 15 N., R. 6 E. at the contact between units 1 and 2.

Specimen 8: volcanic sandstone matrix of breccia from SW¼ sec. 23, T. 15 N., R. 6 E. near lower contact of unit 2 (characteristic of volcaniclastics of unit), medium gray green, well indurated, fairly well sorted, with medium- to fine-grained fragments.

Rounded to angular fragments set in 15 per cent argillaceous matrix containing small feldspar laths and glass fragments altered to zeolite. Fragments submicroscopic to 0.8 mm in thin section. Thirty per cent plagioclase grains (An_{24} to An_{30}, fragments of larger crystals, most shot through with secondary calcite and sericite), 15 per cent chlorite (mainly altered from pyroxene, fills interstices around rock fragments and within fractures of plagioclase), 5 per cent magnetite (disseminated grains and alteration rims around chlorite, mostly altered to hematite), 5 per cent calcite (altered from plagioclase), 30 per cent andesite rock fragments (about 5 types with similar compositions but different textures, colors, and percentages of phenocrysts; some have matrices altered to zeolite)

UNIT 3: Conformably above unit 2 is about 400 feet of sandstone and shale which apparently pinches out southward. The rocks are similar to those of unit 1: massive and well bedded, well to fairly well sorted, medium- to fine-grained, light- to dark-gray and brown muscovite-bearing sandstones interbedded with gray to black and brown, massive to fissile shales. Some of the sandstones show cross lam-

inations, and one outcrop of black, well-indurated shale shows spheroidal weathering. Carbonaceous fragments and poorly preserved leaf fragments are found along bedding planes in the shales and fine-grained sandstones.

Near the upper contact, the sandstone is fine-grained and dark gray and is interbedded with black shale. This upward change is similar to that of unit 1. Above the upper contact of unit 3, black friable shale is interbedded(?) with volcanic conglomerate and breccia, but upward the section becomes entirely volcanic.

Specimen 13: well-indurated, light- to medium-gray, coarse- to medium-grained, fairly well sorted, muscovite-bearing arkose from massive sandstone outcrop with abundant carbonaceous fragments, S½ NE¼ sec. 25, T. 15 N., R. 6 E.

Thirty-five per cent angular quartz grains (some strained, some derived from vein quartz), 20 per cent plagioclase (An_{29} to An_{35}, angular, somewhat kaolinized), 5 per cent potassium feldspar, 15 per cent biotite, 4 per cent muscovite (both biotite and muscovite bent around resistant grains), 10 per cent chert and quartzite fragments, 5 per cent secondary calcite, 1 per cent leucoxene, <1 per cent chlorite (alteration product of biotite), 5 per cent matrix (argillaceous, stained by iron oxide). Two sandstone granules and a volcanic fragment observed. Mineral grains average 0.06 mm, range from 0.08 mm to submicroscopic.

UNIT 4: Rocks of unit 4 are volcanic conglomerate, breccia, and sandstone, mostly gray and gray green, weathering to medium brown to buff. The lowest part of the sequence is volcanic-pebble conglomerate succeeded upward by volcanic breccias, conglomerates, and sandstones. At the top of the unit, medium- to fine-grained well-bedded volcanic sandstone is in contact with muscovite-bearing sandstone and shale of the overlying unit.

Medium-gray porphyritic andesite was observed near the lower contact. Near the top of the sequence, a light gray-green andesite with a few milky plagioclase phenocrysts as much as 3 mm long and a very fine-grained black basalt crop out.

Specimen 14: volcanic-pebble conglomerate from NE¼ NE¼ sec. 25, T. 15 N., R. 6 E. near lower contact of unit 4, well-indurated, gray brown to olive greenish brown with scattered carbonized wood fragments; most grains angular to subangular, are olive-green, gray, and black fragments of porphyritic andesite.

Fifteen per cent matrix (dirty-brown, argillaceous, with submicroscopic particles and secondary calcite grains), 85 per cent rock and mineral fragments: 1 per cent angular quartz, 3 per cent carbonized wood fragments, 20 per cent untwinned plagioclase. Most rock fragments altered to calcite. Some andesite fragments very well rounded, most angular to subangular; many altered to chlorite, some unaltered

UNIT 5: This unit crops out in the Copper Creek drainage basin with an outcrop width of about 2 miles because of repetition in a minor anticline and syncline. The upper and lower contacts diverge slightly to the north, suggesting thickening of the unit in that direction, although widening of the outcrop area may be entirely due to folding. Volcanic rocks above this unit appear to be conformable with it; their contact extends continuously northward to the Puyallup River (Willis, 1886, geologic map).

Lithologically this unit is, like units 1 and 3, a rather monotonous sequence of muscovite-bearing, gray and brown, well-sorted sandstone and arkose interbedded with gray and black shale and thin coal layers. Leaf fossils were collected in two localities near the middle of the unit.

Unit 5 is also exposed south of the Nisqually River. In SE¼ sec. 32, T. 15 N., R. 7 E. and NE¼ sec. 4, T. 14 N., R. 7 E., arkose with some interbedded shale is underlain by tough, gray-green and green andesite flows which may be equivalent to volcanic sandstone of the upper part of unit 4 at SW¼ sec. 19, T. 15 N., R. 7 E. The arkose is light gray, medium- to coarse-grained, well sorted, and fairly well indurated with conspicuous muscovite flakes and occasional thin carbonaceous bands.

Specimen 18: light-gray, well-sorted, fine-grained, fairly well indurated arkose from S½ sec. 6, T. 14 N., R. 7 E. about half way up north slope of Osborne Mountain.

Thirty per cent quartz (most grains strained, one exhibits metamorphic-pavement texture), 20 per cent plagioclase (grains range from An_{26} to An_{34}, some have bent twinning planes), 2 per cent potassium feldspar (some microcline), 15 per cent biotite (flakes fairly fresh, some altered to bleached tan variety with weak pleochroism; flakes have been mechanically squeezed into cracks within some minerals and around resistant grains, in some places are slightly granulated), 5 per cent chlorite (altered from biotite), 5 per cent muscovite (bent, but less so than the biotite), 10 per cent calcite (secondary, in part altered from plagioclase), 15 per cent matrix (cherty appearance).

Mineral grains average about 0.2 mm; most angular, some subangular

The author collected and Roland Brown of the U. S. Geological Survey identified fossil plants from unit 5 in the Copper Creek drainage basin. They appear to be from near the

middle of the unit. The fossils from University of Washington locality WA 231 (NE¼ sec. 19, T. 15 N., R. 7 E.) include *Carya magnifica* (Knowlton) La Motte, *Cercidiphyllum arcticum* (Heer) Brown, *Glyptostrobus dakotensis* Brown, and *Ulmus* sp. Those from University of Washington locality WA 232 (NE¼ sec. 18, T. 15 N., R. 7 E.) include *Cercidiphyllum arcticum* (Heer) Brown and *Carya magnifica* (Knowlton) La Motte.

According to Brown (personal communication), the fossils from Copper Creek are Eocene, "perhaps late." *Carya*, *Ulmus*, and *Metasequoia* "could be regarded as streamside or lowland inhabitants", and *Glyptostrobus* may indicate "a moist environment"; definite statements are difficult without large collections and more species.

The rocks lie at the southern end of a belt of sedimentary rocks that Willis (1886; 1898) described in detail; Willis called these rocks the "Coal Measures" and gave their tentative date as "Laramie". His geologic map (1886) extends almost to the Nisqually River.

White (1888) first used the name Puget group for the rocks described by Willis and, although he did not visit the region, he considered the group to be Eocene from the relations established by Newberry and Willis.

Willis (1898), using the term Puget group, described the rocks in more detail, and subdivided them into the Pittsburgh (youngest), Wilkeson, and Carbonado formations. He later changed the name Pittsburgh formation to Burnett formation.

Culver (1919), using the term Puget series, briefly described the rocks in connection with a coal survey and contributed additional geologic maps of rocks similar to the Puget group as described by Willis, but his correlations were based only on lithology.

Weaver (1937) held that the major portion of the Puget group was deposited in middle and late Eocene time.

Their fossil content, coal-bearing character, and general lithology suggest that the sedimentary rocks within the Nisqually area are probably the equivalent of the Puget group. However, it is inadvisable to assign the sedimentary rocks definitely to the Puget group because the area to the north, which connects the present area to the type area of the Puget group, has not been mapped, and because the two intertongues of volcaniclastic rocks are present in this section of rocks but are not present in most sections of the Puget group.

Volcanic Rocks

Conformably above the sedimentary rocks lies a thick and complex series of volcanic flows, breccias, conglomerates, sandstones, shales, and pyroclastic rocks. The flows, and recognizable rock fragments within the volcaniclastic rocks, are dominantly andesitic, although basalt, rhyolite, and dacite occur. Structure sections indicate a possible thickness of 20,000 feet, but undiscovered faults may have caused repetition of parts of the section, so that the thickness may be no more than half this.

For convenience, these rocks are subdivided into three lithologic units (the lower, middle, and upper parts). It is doubtful that these subdivisions can be traced far beyond the present area.

CONTACT BETWEEN THE SEDIMENTARY AND VOLCANIC SERIES: The lower part of the andesitic series lies with apparent conformity on the sedimentary series. North of the Nisqually River the beds in both series strike slightly west of north and dip about 20° E. The line of contact extends 8 miles north of the area to the Puyallup River, and the same (presumed) contact reappears 5 miles north of the Puyallup River along Mowich Lake road.

The lowest observed rocks in the volcanic series are volcaniclastics. In SE¼ sec. 17, T. 15 N., R. 7 E., crudely bedded, well indurated, and very poorly sorted volcanic breccia lies above the sedimentary series. About 3 miles to the southeast in NE¼ sec. 33, T. 15 N., R. 7 E., along State Highway 5, is a similar volcanic breccia. Its fragments are angular, subangular and rounded, dark-gray to black dense basalt(?) and olive-green and gray porphyritic rocks averaging about 25 mm in longest dimension, although they vary greatly in size. The rock has no visible bedding planes or laminae and is very poorly sorted.

South of the Nisqually River the volcanic-sedimentary contact was crossed on Osborne Mountain in SW¼ sec. 7, T. 14 N., R. 7 E., where coarse-grained volcanic sandstone lies above muscovite-bearing arkose. Stratigraphically above the sandstone is volcanic breccia similar to the breccia north of the Nisqually River.

Flow rocks, pyroclastic rocks, and epiclastic volcanic rocks higher in the section indicate a long and continuing history of volcanic eruption concomittant with erosion. The presence

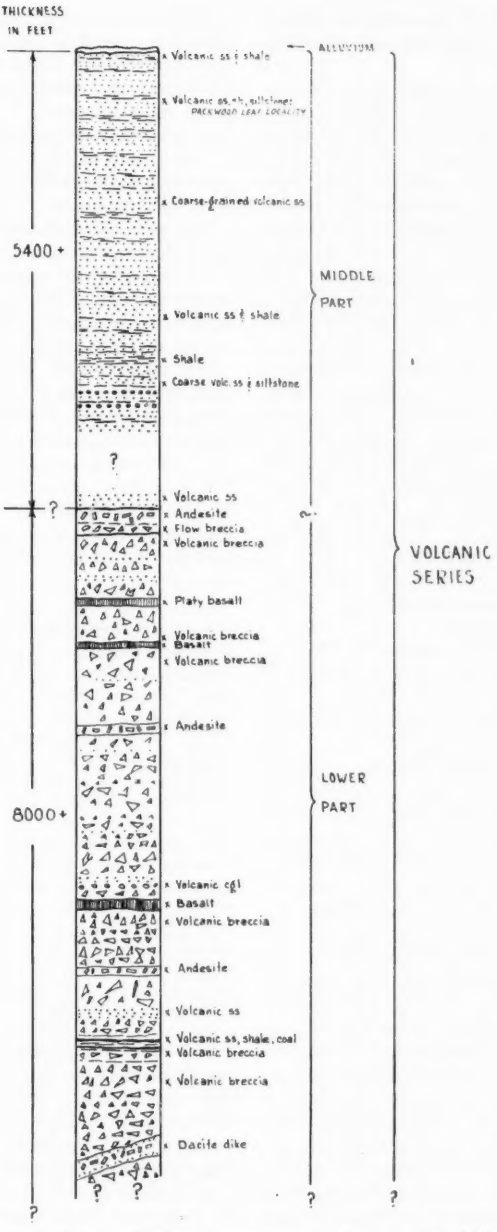


Figure 3. Generalized columnar section showing the middle and lower parts of the volcanic series in the vicinity of Willame Creek. Measurements taken from structure section. X indicates specimen localities

of flows higher in the section indicates that as time progressed the centers of volcanic action became more widespread.

LOWER PART: The lower part of the volcanic series includes a great variety of rock types, including volcanic breccias of laharic, pyroclastic, and probable flow origin, tuff, lapillistone, tuffaceous volcanic sandstone, epiclastic volcanic sandstone and conglomerate, volcanic siltstone and shale, and andesite, basalt, and dacite flows. The rocks vary widely in color but are prevailingly green and gray green. The lower part of the series, as measured from a structure section, is more than 8000 feet thick (Fig. 3).

Fossil leaves were collected about 5500 feet stratigraphically above the sedimentary series exposed on Osborne Mountain by Roland Brown, who dates them as "apparently late Eocene" (Personal communication). Fossil leaves collected from this locality, U. S. Geological Survey Catt Creek locality (SE¼ NE¼ sec. 13, T. 14 N., R. 6 E.), include *Lastrea fischeri* Heer, *Metasequoia occidentalis* (Newberry) Chaney, *Carya* sp., *Salix* sp., *Comptonia* sp., *Betula* sp., *Alnus* sp., *Calyptranthes* sp., and *Persea* sp.

The volcanic breccias are generally massive, usually chaotically sorted, and commonly well lithified. Some beds are as much as 100 feet thick in single outcrop without visible bedding, as along the northern side of Sawtooth Ridge where glaciers have carved precipitous cirque walls. Here, bedding planes in the massive breccias can be distinguished only at a distance, because the expression of bedding is change in color. The breccias are interbedded with andesite flows, volcanic sandstone, shale, and tuff beds. Some breccias contain disseminated carbonaceous fragments.

Rock fragments of the breccias are aphanitic and porphyritic, although some are vesicular and amygdaloidal. These fragments are various shades of red, gray, and green; some are black. They are angular to subangular to well rounded and average about 2 inches in longest dimension, although some are as much as 6 feet in diameter. Four to five distinguishable rock types commonly occur within a single breccia deposit, but some deposits have only one or two rock types. In places, there are unoriented carbonized wood fragments. In some breccias the fragments blend (grade?) into a usually lighter green-gray matrix, but most fragments have sharp outlines. The breccias in places grade laterally and/or vertically into volcanic conglomerates or sandstones.

The matrix of the volcanic breccias, composed of mineral grains and small rock chips similar to the large fragments, varies from light green gray to medium green and black. The green results mainly from secondary chloritic material. Milky-white plagioclase grains consisting mainly of broken fragments of larger crystals are ubiquitous. Augite is present in many of the rocks, and hornblende is rare.

The most abundant type of volcanic breccia in the lower part of the volcanic series is apparently of epiclastic origin. Pyroclastic breccias are less abundant. The writer (1960) has given criteria to distinguish these breccias and described two breccias of this unit in detail.

Volcanic sandstones closely resemble the volcanic breccias except for grain size. They are typically green when fresh but weather to light brown of various shades and are usually very well indurated and occasionally siliceous(?). The sandstones are commonly interbedded with olive-green siltstones and/or pebble conglomerates, but in some places are interbedded with greenish tuff, lapillistone, and pyroclastic breccia.

Specimen 51: dark-green, well-indurated, well-sorted, coarse- to medium-grained volcanic sandstone with angular and some round grains from W½ projected section 4, T. 13 N., R. 8 E., from western end of Willame Creek about 1 mile northwest of Long Lake.

Eighty five per cent rock and mineral fragments (angular to rounded), 10 per cent matrix (very fine-grained argillaceous material, much of which is altered to chlorite). Three varieties of andesite fragments (most have felty textures and devitrified glass matrices, a few are vesicular, some are altered to hematite). Most plagioclase (An_{26}, An_{28}, An_{30}) angular, fairly fresh; some altered to kaolin and partially to calcite. Leuxoxene, magnetite, and augite(?) less than 1 per cent. Chlorite occurs throughout matrix, in volcanic fragments, within feldspar grains (altered ferromagnesian inclusions). Secondary calcite (<1 per cent) present within the matrix and plagioclase

The volcanic conglomerates, typically poorly consolidated (some are well lithified), are waterlaid volcanic materials with well-rounded, sand- to boulder-sized fragments. They commonly weather yellow brown to dark brown. The rock fragments of some layers are coated with black to blue-black manganese oxide and are composed of andesite or basalt set in a matrix of tuffaceous(?) volcanic sandstone. Megascopically the large fragments appear to be basalt, but microscopically they are identified as andesite. Carbonaceous fragments are common within these deposits. The conglom-

erates are usually associated with dirty-brown to dark-gray, poorly to well-indurated siltstones and sandstones.

Some conglomerates are more closely related to the volcanic breccias in color, fragment type, and induration, and all gradations exist between breccia and conglomerate.

The heavy forest cover makes correlation of the many outcrops of volcanic conglomerate in the mapped area difficult. Moreover, conglomerate apparently recurs vertically within the volcanic sequence.

Shale is interbedded with volcanic sandstone, conglomerate, and breccia and in places is in contact with lava flows. The shale is invariably silty but is also sandy and in places has scattered pebbles or pebble and sandstone lenses. Colors are light to dark gray and black. Green and red shales are minor. Beds of the shale range from massive to paper thin and may be fissile or platy. Conchoidally fractured shale also occurs. Fragments of carbonized wood are present nearly everywhere, and leaf imprints are found along many bedding planes. Thin coal units are interbedded in places with the shale.

Volcanic flows occur widely throughout the mapped area. The upper and lower contacts of single flows are rarely observed in the same outcrop, and usually neither contact can be seen. The thicknesses of most individual flows are therefore unknown but apparently range from a few feet to several tens of feet. Individual flows are in places interbedded with volcaniclastic rocks.

The majority of flows are structureless, but some show crude columnar jointing, platy structures, or crude flow banding. Flows with well-developed columnar jointing and platy structures are generally dark gray to black and may be younger than the typically altered green flow rocks.

The flows vary widely in megascopic appearance. In the field they are identified as andesite, basalt, dacite, and rhyolite in decreasing order of abundance. The rhyolite and basalt are generally aphanitic, whereas the andesite and dacite are porphyritic, usually with abundant plagioclase crystals, commonly with megascopic pyroxene and less commonly with amphibole crystals. Colors are white, black, grays, browns, greens, and reds.

The most abundant flows are porphyritic andesites. Phenocrysts are plagioclase and less commonly augite crystals. Hornblende is conspicuous in a few rocks but is generally absent. The most characteristic flows are green porphyritic andesites with fresh to altered feldspar and augite crystals and abundant secondary chlorite.

Specimen 43: gray to gray-green augite andesite with aphanitic matrix and abundant large anhedral to euhedral plagioclase and pyroxene phenocrysts from a sequence of sandstone and tuffaceous(?) sandstone, from NW¼ projected section 11, T. 13 N., R 8 E. along Willame Creek

Matrix is small (average 0.08-mm) sutured mineral grains, mainly plagioclase, pyroxene, and possible quartz and zeolite, and abundant small dust-sized magnetite(?) grains. Plagioclase and augite phenocrysts euhedral, up to 3 mm

Twenty per cent plagioclase phenocrysts (An_{45}, show kaolinitic alteration along very fine feather-like fractures), 35 per cent matrix plagioclase crystals (untwinned, irregularly shaped), 20 per cent augite grains (mainly unaltered, some with borders altered to magnetite and chlorite, some completely altered to chlorite and magnetite; most twinned and commonly euhedral), 10 per cent chlorite (green fibrous variety with slight pleochroism, altered from matrix material and some pyroxene phenocrysts), 5 per cent magnetite (as small discrete grains throughout matrix, and as partial alteration of pyroxene), <1 per cent quartz crystals (irregularly shaped, may be secondary), 5 per cent kaolin (altered from plagioclase). About 5 per cent of matrix consists of undetermined fine-grained material.

MIDDLE PART: The middle part of the volcanic series consists of more than 5400 feet of east-dipping volcanic breccias, conglomerates, sandstones, and shales along the eastern border of the area near Skate Creek. The lower contact may be conformable with the underlying part of the series, but no actual contacts were seen. The middle part disconformably underlies the upper part of the series and may thin or interfinger with volcanic rocks toward the west, for no comparably thick recognizable series of volcaniclastics was found to the west. These rocks are overlapped northward by the overlying upper unit (interpretation based upon map pattern).

Most of the rocks are greenish-gray, but some are gray and brown. The brown is due mainly to weathering of the matrix material. Black carbonaceous shales are interbedded with rocks of this unit in places.

Fossils collected from near the top of this unit indicate an Eocene age. The following were collected by the writer and separately by Roland Brown, and all fossils were identified by Brown. From University of Washington locality WA 235 (SW¼ sec. 8, T. 13 N., R. 9 E., near Packwood along Skate Creek): *Castanea castaneaefolia* (Unger) Knowlton, *Metasequoia*

occidentalis (Newberry) Chaney, *Glyptostrobus dakotensis* Brown, and *Laurus* sp. From U. S. Geological Survey Packwood locality (same location): *Adiantum anastomosum* Brown, *Cercidiphyllum elongatum* Brown, *Dillenites* sp., *Metasequoia occidentalis* (Newberry) Chaney, *Platanus* sp., and *Sequoia affinis* Lesquereux.

According to Brown (Personal communication), these specimens are similar to those at Steel's Crossing (sec. 11, T. 13 N., R. 4 E., about 2 miles west of Renton, Washington) dated as late Eocene. Significantly, the rocks in which the leaves occur are apparently several hundreds of feet stratigraphically above the sedimentary series of the Nisqually River valley.

Specimen 55: dark-green, poorly sorted, very coarse-grained volcanic sandstone with scattered rounded andesite pebbles as much as 7 mm in long dimension from NW¼ sec. 18, T. 13 N., R. 9 E. along a tributary of Willame Creek

Eighty per cent fragments, 20 per cent matrix. Rock fragments (50 per cent) of about six different types. Largest and most numerous are porphyritic andesites (one very light in color, with abundant magnetite dust and matrix of ragged feldspar laths and devitrified glass; plagioclase phenocrysts of this fragment (An_{35}) relatively fresh). A few of the smaller rock fragments are rounded, have dirty-brown matrices with pilotaxitic textures. Other fragments are irregular in shape, reddish, have felty textures, and have more calcic feldspar phenocrysts (An_{45}). Twenty per cent plagioclase grains, An_{27} and An_{47}–An_{60} (mostly fresh, some slightly altered to kaolin; most angular, some have sharp euhedral outlines). Less than 1 per cent each unaltered pyroxene grains, leucoxene, magnetite. One epidote grain observed. Chlorite (10 per cent) fills most interstices.

UPPER PART: The upper unit includes tuff, lapillistone, volcanic sandstone, siltstone, and volcanic breccia. Volcanic flows in the Silver Creek depression and along the eastern border of the area are also included. This unit may be as much as 5000 feet thick.

Leaves collected by Roland Brown from tuff near the base of the unit (in S½ sec. 19, T. 14 N., R. 7 E.) are tentatively dated as Oligocene. The regional structural pattern suggests that volcanic sandstone and shale (tuffaceous?) found in secs. 13, 14, 22, and 23 in the Silver Creek region may be equivalent (Fig. 4).

Leaves collected from the U. S. Geological Survey "Creaking Tree" locality (SE¼ sec. 19, T. 14 N., R. 7 E. include *Metasequoia occidentalis* (Newberry) Chaney, *Cercidiphyllum* sp., *Betula* sp., *Persea* sp., *Ulmus* sp., and *Acer glabroides* Brown.

The rocks at the "Creaking Tree" locality are light-colored (gray and tan) tuffaceous rocks lithologically unlike the greenish, altered flows and volcaniclastics of the underlying series. Because their structural attitude diverges slightly from the underlying rocks, they appear to be unconformable; the contact was not seen. Brown states (Personal communication), "The presence of *Acer glabroides* in this collection, as well as the general composition of the collection, is suggestive of Oligocene age."

Specimen 41: light-tan, fine-grained tuff with light-colored rock and mineral fragments and carbonaceous matter, all of which are roughly parallel to the bedding planes.

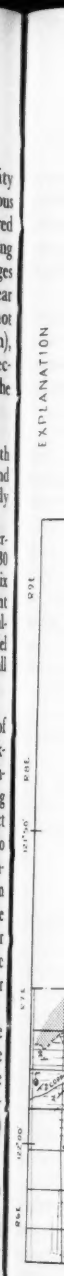
Sixty per cent matrix (composed of undetermined submicroscopic, birefringent fragments), 30 per cent calcite (irregularly shaped, includes matrix material; some altered from plagioclase), 9 per cent angular glass fragments (some partially recrystallized, mainly elongate, oriented roughly parallel to each other), and less than 1 per cent small angular quartz fragments.

Volcanic flows along the eastern border of the area total are more than 2000 feet in thickness. The map pattern suggests that the contact between the flows and the underlying units is disconformable, although the contact was not observed. The rocks are dark-gray to light-brown and gray, crudely columnar andesite flows. They crop out extensively from near Johnson Creek to the mouth of Skate Creek and may extend into Mount Rainier National Park to the north. Near Willame Creek, columnar andesites form a spectacular 500–600-foot cliff.

This series of flows is included within the upper unit because structure sections across the area suggest their equivalence to flows in the Silver Creek depression which lie stratigraphically above the light-colored Oligocene(?) tuffs exposed near Catt Creek.

Specimen 53: reddish-gray to reddish-brown porphyritic andesite from W½ projected sec. 22, T. 14 N., R. 8 E. near junction of Johnson and Skate creeks. In outcrop, flow shows crude columnar jointing and blocky jointing; jointing near center of flow is platy.

Groundmass composed of weakly birefringent irregularly shaped minerals (zeolite?) and abundant small lath-shaped feldspar crystals and small ferromagnesian minerals (augite?) altered to antigorite and chlorite. Plagioclase phenocrysts (An_{30}) appear ragged; many partially altered to sericite and kaolin. Augite in various stages of alteration to antigorite and chlorite; some completely altered. Magnetite present as scattered specks.



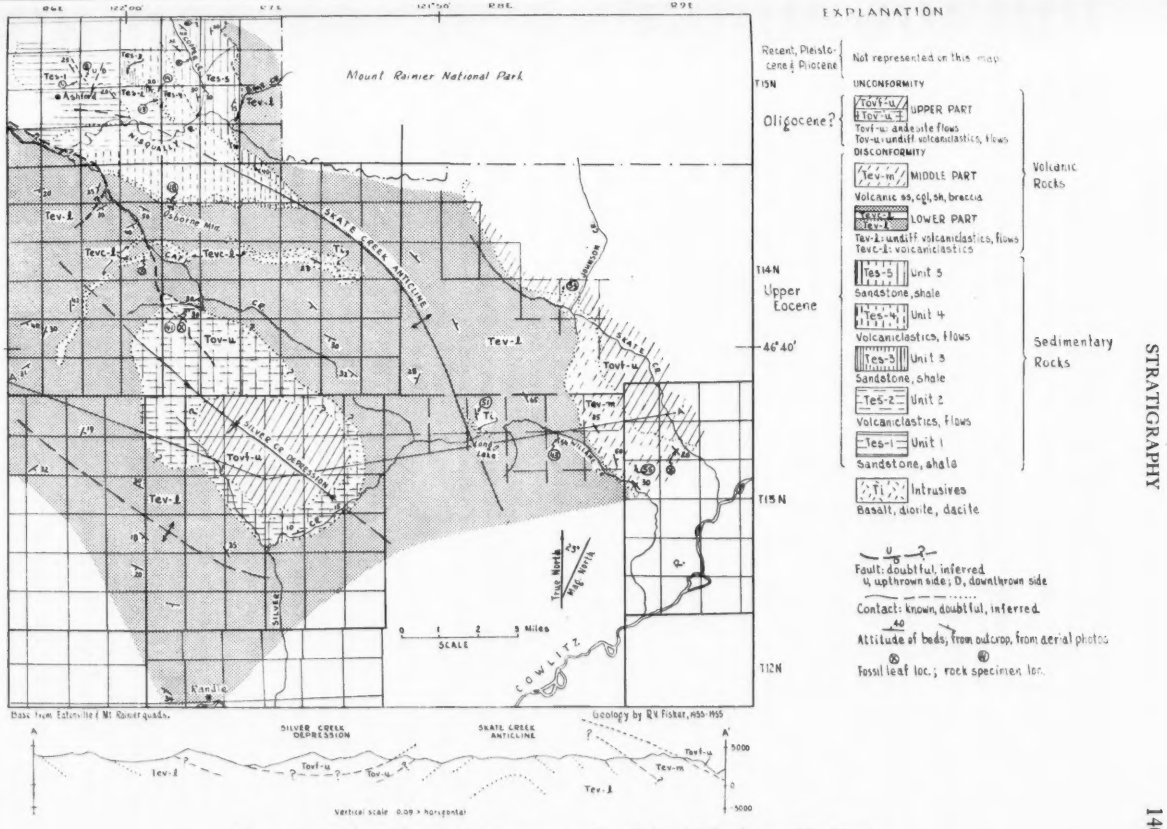


Figure 4. Geologic map and structure section of the Ashford area, Washington

AGE AND CORRELATION: Fossil leaves indicate that the volcanic sequence is upper(?) Eocene to Oligocene(?), but the evidence is too meager for definite dating.

The rocks of the volcanic series are lithologically similar to parts of the Keechelus andesitic series as described by Smith and Calkins (1906), Coombs (1936), and Warren (1941), and perhaps could be assigned to that formation. However, the original description of the Keechelus formation by Smith and Calkins did not include a type section with well-defined upper or lower limits, nor were the age assignments certain. And, although Coombs (1936) used the designation "Keechelus andesitic series" for basement rocks of Mount Rainier in the area adjacent to the present area, the age determinations of the volcanic series in the present area are too uncertain, and the stratigraphic succession cannot be matched in any way to the originally described Keechelus formation. Also, the rock types and stratigraphic succession do not correspond to the new type area proposed for the Keechelus by Foster (1960, p. 120).

In view of the uncertainties of correlation, it seems best to await more detailed work before assigning formational names to the volcanic rocks described herein.

CONDITIONS OF SEDIMENTATION

The rocks of the area, primarily terrestrial in origin, represent a period of relatively rapid sedimentation during late(?) Eocene time. Materials of the earliest phase of sedimentation were derived from plutonic and metamorphic sources, probably from an easterly direction (Weaver, 1937, Pl. 2). Well-bedded sandstone (with translation ripple marks and cross laminations) and shale, thin coal units, and abundant leaf imprints in the lower part of the sequence indicate a region of low relief probably near sea level. Andesitic volcanism, recorded by volcanic units 2 and 4 intertongued with nonvolcanic sedimentary rocks, began in late(?) Eocene time. The presence of coarse volcaniclastic rocks and volcanic flows shows that the volcanoes were not far distant. Following this early period of volcanism and erosion (which may have continued, but with the products carried elsewhere), more nonvolcanic sediments were deposited, followed by a period of andesitic volcanism which blocked the source of nonvolcanic sediments. Rocks of the nonvolcanic and volcaniclastic sequence show no recognizable mixing of plutonic-metamorphic minerals with volcanic materials.

Volcanism following the deposition of nonvolcanic sediments has been persistent through Tertiary and Quaternary time. The abundance of thick well-bedded epiclastic volcanic rocks —including volcanic sandstones and shales which commonly contain abundant leaf imprints and occasional thin coal layers—, the interbedding of these epiclastic rocks with flows and pyroclastic deposits, and the absence of marked angular unconformities all indicate a region of fairly low relief receiving stream transported and air-borne products of volcanoes deposited in lakes, streams, and valleys, concommittant with erosion and at times thick outpourings of lava.

The combined thickness of the volcanic and sedimentary formations in the area, the abundance of well-bedded epiclastic volcanic sandstone and shale, and the absence of marked angular unconformities indicate that the area was subsiding and that the blocking of the source for nonvolcanic materials was probably due to the construction of volcanoes rather than to uplift.

STRUCTURAL GEOLOGY

The structural pattern of the mapped region is one of broad and open, northwest-trending folds that are oblique to the north-south trending Cascade Mountains. This pattern is dominated by a large anticline (Skate Creek anticline) near the eastern boundary of the area and a broadly folded structural low (Silver Creek depression) to the west. Smaller and more local folds within the rocks of the area are most easily recognized in the well-bedded and less competent sedimentary series. Faults are difficult to recognize but seem to be of minor importance except for the northwest-trending Catt Creek fault. Time of deformation is not known, although the rocks were undoubtedly involved in the late Miocene period of folding in Washington and Oregon (Weaver, 1937, p. 198).

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Short Notes

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PROPOSED CLASSIFICATION OF VOLCANICLASTIC SEDIMENTS AND ROCKS

Abstract: Volcaniclastic sediments and rocks are divided here into autoclastic, pyroclastic, and epiclastic types with grain-size limits the same as nonvolcanic epiclastic rocks. Autoclastic rocks contain fragments that are produced within (but not usually extruded from) volcanic vents, during movement of lava flows, or by gas explosions within flows that have ceased to flow. Pyroclastic rocks contain fragments produced by volcanic explosion and extruded as discrete particles from volcanic vents. Epiclastic volcanic rocks contain fragments produced by weathering and erosion of solidified or lithified volcanic rocks of any type. Volcaniclastic types may be mixed in all proportions with each other or with nonvolcanic fragments, although these mixtures are not designated within this classification. A nongenetic category, based only upon particle size and the presence of volcanic material, is included for rocks with clasts of unknown origin.

Many classifications of pyroclastic rocks and/or pyroclastic fragments have been suggested in the past and are reviewed by Wentworth and Williams (1932). Only recently, however, have attempts been made to group all volcanic clastic (termed *volcaniclastic* in this paper) rocks into a single system (Blokhina *et al.*, 1959; Pantó, 1959). There is a need for such comprehensive systems, because rocks of pyroclastic origin are only one category of volcaniclastic rocks. "Pyroclastic" is clearly defined and widely used as "an adjective applied to rocks produced by explosive or aerial ejection of material from a volcanic vent. . . ." (Wentworth and Williams, 1932, p. 24–25). If this definition is to be accepted and used, the following cannot be considered pyroclastic: (1) fragments produced in volcanic vents but not ejected from the vent; (2) fragments formed during the breaking up of a moving lava flow; (3) fragments that originate by the weathering and erosion of solidified lava flows or consolidated pyroclastic deposits.

A simple working classification that incorporates all the factors involved in the development of volcaniclastic rocks is clearly impossible. The one developed here is based on (1) the primary origin of the fragments, and (2) their grain size. In addition, a purely descriptive category is included for the volcaniclastic rocks of doubtful or unknown origin.

There is complete gradation between nonvolcanic sedimentary rocks and extrusive igneous rocks, and all types may be included within the volcaniclastic rocks. The pigeonholing of all types into separate categories is not possible, nor is it advisable, because our knowledge of the volcaniclastic rocks is somewhat limited by our ability to recognize their various origins. Criteria of recognition have been given by Pirsson (1915) on tuff, by C. S. Ross (1928) on reworked pyroclastic and/or epiclastic volcanic rocks, and by Enlows (1955) and R. L. Smith (1960) on welded tuff. A superb contribution to our knowledge of marine volcaniclastic rocks has been made by W. R. Dickinson (1958, Geology of the Izee area, Grant County, Oregon, unpublished Ph.D. thesis, Stanford University). Discussion of criteria for recognizing the types of volcaniclastic rocks is beyond the scope of the present article.

Grabau (1924), in his masterful although largely unaccepted scheme of rock classification, has in part solved the problem. His exogenetic (clastic) subdivision of rocks includes pyroclastic as one of five major types. Pyroclastic is further subdivided by grain size using the terms rudyte (>2.5 mm), arenyte (0.5 mm–2.5 mm), and lutyte (<0.5 mm). These same terms (with "ite" used instead of "yte") are in use today (Dunbar and Rodgers, 1957), but with Grabau's size limits changed to agree with the Wentworth scale of grade sizes. Grabau's pyrorudyte is a pyroclastic rock cor-

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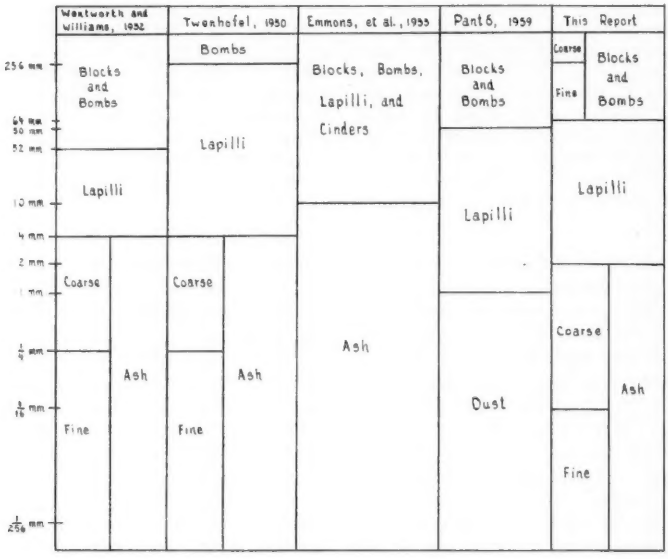
responding in grain size to other rudytes (conglomerate and breccia), pyrarenyte corresponds in grain size to other arenytes (sandstone), and pyrolutyte corresponds in grain size to other lutytes (shale). He thus equated pyroclastic grade sizes to epiclastic grade sizes. Furthermore, he clearly recognized that pyroclastic rocks may be reworked by erosional agents and prefer indefinite comparisons of pyroclastic debris to pea and walnut sizes.

The limits of pyroclastic particles should be reset because quantitative size analyses of pyroclastic debris are made by sedimentary petrographic methods, because pyroclastic size limits in use are not uniform, and, most importantly, because such a change would

TABLE 1. CLASSIFICATIONS OF PYROCLASTIC FRAGMENTS



prefixed hydro (water), anemo (wind), and atmo (weathering), giving hydropyrorudyte, anemopyrarenyte, etc.

In order to include all volcaniclastic rocks within a single system, it is necessary to use similar limiting grade sizes. Studies of pyroclastic debris by Moore (1934), Baak (1949), Segerstrom (1950), Thorarinsson (1954), and others show the tendency for use of standard sediment-size terms. Baak uses the U. S. Bureau of Soils scale, Thorarinsson uses the Atterburg scale, Segerstrom and Moore use the Wentworth scale.

Thorarinsson (1954, p. 1) points out that size groups of pyroclastic fragments are not precisely defined. This is not strictly so, but defined size limits of pyroclastic fragments vary (Table 1). Some authors (Holmes, 1946, p. 451; Stearns and Macdonald, 1946, p. 16) still relate pyroclastic, epiclastic, and pyroclastic-epiclastic mixtures within a single scheme. The size scale used here (Table 2) is the Wentworth scale as recommended by Lane *et al.* (1947), except the word "pebble" is substituted for "gravel."

Blocks and bombs are here equated to sediment terminology by setting size limits of coarse blocks and bombs to correspond with boulder size (>256 mm), fine blocks and bombs to correspond with cobble sizes (64 mm–256 mm), and lapilli to correspond with pebble sizes (2 mm–64 mm). Coarse ash is equivalent to sand sizes (2 mm–1/16 mm), and fine ash is equivalent to silt- and clay sizes. The wide range given to fine ash is more apparent than real because ash of clay size is volumetrically unimportant (Baak, 1949, p. 54–55).

Volcanic rocks are herein divided first on the basis of origin of particles (autoclastic, pyroclastic, epiclastic), and secondarily on particle size (Table 3). A nongenetic category is included based only upon presence of volcanic particles and their sizes. Secondary classifications using different criteria may be set up under the proposed classification. For example, material of "fine-dust" size (Macdonald, 1953). Gas entrapped in lava that has ceased to flow or in pyroclastic flows after deposition may continue expansion and disruption of solidified or partially solidified material, thus producing clastic zones (Earl F. Cook, written communication, 1961). Processes within volcanic pipes may also produce fine- and coarse-grained frag-

TABLE 2. PROPOSED TERMINOLOGY AND GRAIN SIZE LIMITS FOR PYROCLASTIC FRAGMENTS

Grade size (mm)	Epiclastic fragments	Pyroclastic fragments	
	Boulders (and "blocks")	Coarse	Blocks and bombs
256			
	Cobble	Fine	
64			
	Pebble	Lapilli	
2			
	Sand	Coarse	Ash
1/16			
	Silt	Fine	
1/256			
	Clay		

further subdivision of tuff may be based upon type of transporting agent (fluvial, glacial, eolian), upon environment of deposition (lacustrine or marine), upon mode of origin of pyroclastic fragments (essential, accidental, accessory), upon composition (crystal, lithic or vitric), etc. The reader is referred to Wentworth and Williams (1932) for a review of various classifications.

Autoclastic applies to all the possible ways fragmental rocks may be produced in volcanic vents, by friction in flowing lava or by internal gas explosions within flows after they have ceased to flow. In flowing lava that is cooling, friction produces large fragments, as well as ments (Hans Cloos, 1941; Reynolds, 1954; Gates, 1959). Tuffisite is in use for clastic material produced in volcanic pipes, and it may be well to restrict the term to ash-sized material. Such clastic material may be composed of any pre-existing rocks derived from the walls of a volcanic pipe as well as clasts derived from the magma. It is possible therefore to have any rock type, including any type of volcaniclastic rock, within the volcanic pipe. In addition, fragmental material produced in the volcanic pipe may flow onto the surface (Curtis, 1954). Hybrid rocks such as these cannot be neatly categorized and should be fully described rather than given a simple name.

Pyroclastic fragments are produced by volcanic explosion and are extruded as discrete particles from vents. "Primary" is used here to indicate that the material has not been moved from its place of deposition before its lithification. For example, "primary tuff" is formed of lithified ash-fall material that has not been reworked by epigene geomorphic agents or by gravity. "Reworked" indicates that ash-fall material has been moved from its original place of deposition and redeposited before lithification, as for example "reworked tuff." Reworked tuff is to be distinguished from rocks formed by epiclastic processes.

Terms for lithified aggregates of bombs and blocks, except for grain-size limits, are similar to Wentworth and Williams' definitions. Pyroclastic breccia is used for lithified aggregates of pyroclastic blocks rather than "volcanic breccia" as proposed by Wentworth and Williams. Agglomerate is a rock aggregate composed mainly of large (>64 mm) pyroclastic fragments rounded by volcanic processes. Coarse and fine denote various size limits of the Wentworth scale. Lapillistone is proposed here for rock aggregates composed predominantly of lapilli (2 mm–64 mm) because there is no term for rocks composed dominantly of lapilli-sized fragments. Lapilli tuff is formed of nearly equal amounts of lapilli and ash and is analogous in use to "tuff breccia." This gives lapilli tuff a somewhat different meaning from that given it by Wentworth and Williams (1932, p. 47). Stevens (1959, p. 119) proposed lowering the size limit of pyroclastic breccia to 2 mm which would eliminate the term lapilli tuff.

The upper size limit for ash in a lithified tuff is changed from 4 mm to 2 mm to agree with sediment terminology. This limit is lowered primarily because situations may arise when one would be forced to decide whether fragments of a rock are pyroclastic or epiclastic in order to name it (*i.e.*, breccia vs. tuff, or breccia vs. sandstone). If the limit of 2 mm is consistent throughout the classification, the nongenetic name volcanic sandstone or volcanic breccia may be used without having to determine the origin of fragments. The size limits for coarse and fine ash (and tuff) are changed from Wentworth and Williams' 1/4 mm to 1/16 mm for consistency. Coarse tuff is equivalent in particle size to sandstone, and fine tuff corresponds to siltstone and claystone. A three-fold subdivision into coarse, medium, and fine to correspond with sandstone, siltstone, and claystone is unnecessary.

Epiclastic is defined as "a term applied to mechanically deposited sediments (gravel, sand, mud) consisting of weathered products of older rocks. Detrital material from pre-existent

TABLE 3. PROPOSED CLASSIFICATION OF VOLCANICLASTIC ROCKS

Predominant grain size (mm)	Autoclastic *	Pyroclastic * Primary or Reworked		Epiclastic * +	Equivalent nongenetic terms * +
256	Flow breccia, Autobreccia, Intrusion breccia	Pyroclastic breccia, Agglomerate		Epiclastic volcanic breccia,	Volcanic breccia,
64		Lapillistone		Epiclastic volcanic conglomerate	Volcanic conglomerate
2	Tuffisite	Coarse	Tuff	Epiclastic volcanic sandstone	Volcanic sandstone
1/16		Fine		Epiclastic volcanic siltstone	Volcanic siltstone
1/256				Epiclastic volcanic claystone	Volcanic claystone

* May be mixed with nonvolcanic clastic material

\+ Add adjective "tuffaceous" to rocks containing pyroclastic material < 2 mm in size

rocks" (American Geological Institute, 1957, p. 97). Epiclastic volcanic rocks are defined as those derived by weathering and erosion of lithified or solidified volcanic rocks. In an earlier paper (1960, p. 978) the author used "epiclastic" in a sense that would include all volcaniclastic material transported by epigene geomorphic agents, including unconsolidated pyroclastic debris. Upon examination of the whole field of volcaniclastic rocks, however, it becomes apparent that rocks containing particles formed by pyroclastic processes and those formed by weathering processes should be, insofar as possible, in separate categories. Comminution of magma or lava by volcanic explosion is unique in the formation of clastic rocks, because liquid is converted directly to solid clastic material. Weathering and erosion accomplishes granulation at a much slower rate, and probably with a greater total expenditure of energy. It is important to distinguish, if possible, fragments produced by explosion from those produced by weathering processes.

Graywacke that contains volcanic fragments (volcanic wackes of Williams, Turner, and Gilbert, 1954) may be included within the epiclastic volcanic rock category, and many are in large part composed of pyroclastic fragments. The bulk of volcanic wackes observed by me in an area of lower Tertiary volcanic rocks in the Cascade Mountains near Mount Rainier, Washington (Fisher, p. 1395, this bulletin), appear to be epiclastic. Pogue (1909), Edwards (1950), and Dickinson (1958, Geology of the Izee area, Grant county, Oregon, unpublished Ph.D. thesis, Stanford Univ.) discuss volcanic wackes formed in probable geosynclinal environments. Williams, Turner, and Gilbert (1954, p. 303–304, 308) give a general account of epiclastic volcanic rocks and graywackes.

Nongenetic names based upon presence of volcanic fragments and minerals and upon their sizes are needed for volcaniclastic rocks with fragments of unknown origin. Volcanic sandstone, for example, is a rock containing volcanic particles of sand size regardless of the origin of the fragments.

Acknowledgments

Many of the author's ideas are the result of correspondence and conversations with several geologists. The instructive hours spent in discussion with Ray E. Wilcox, on the field of volcanic rocks in general and pyroclastic rocks in particular have greatly aided the writer in formulating his ideas. Comments and suggestions on the classification by Richard L. Hay, and George I. Smith are appreciated. The writer wishes to thank Howel Williams, Julian D. Barksdale, and Robert M. Norris for valuable suggestions on the original manuscript, and Betty Skipp and Earl F. Cook for suggestions on the present manuscript. Discussion and correspondence with William R. Dickinson has been very helpful. Responsibility for the classification, of course, rests with the author.

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DÉCOLLEMENT STRUCTURES IN GLACIAL-LAKE SEDIMENTS

Abstract: A series of unconsolidated Pleistocene sediments displays décollement folds and faults on a small scale. Correlation between directions of ice movement inferred from till fabrics and the spatial arrangement and shear sense of folds in underlying lake clays indicates that deformation was induced by overriding glacier ice. Fold axes produced by this single deformative event are seen in many orientations relative to the direction of movement of the glacier; some axes are essentially parallel to direction of the glacier movement.

At the southeastern margin of the Adirondack Mountains, 5 miles southwest of Glens Falls, New York, three road cuts (outcrops 1, 2, 3) on the north side of the principal unnumbered route along the north bank of the Hudson River exhibit glacial-lake sediments overlain by stony till (Fig. 1). Where undisturbed, the lake sediments are evenly laminated and laterally persistent. The sediments display rhythmic lamination; dark clay-rich layers alternate with light-colored more silty layers. The dark layers are medium gray (N5; color designations, according to the Munsell Rock Color Chart, are for air-dried samples) and consist mainly of clay with about 10 per cent silt and fine sand. The lighter layers are yellowish gray (5Y 7/2) and contain more silt and fine sand (20 per cent). Several layers of moderately well sorted, medium- to very-fine sand are interbedded with the finer sediments. Boulders, cobbles, and pebbles are plentiful and are probably the result of ice rafting. Boulders range up to 3 feet in longest exposed dimension. The lake sediments immediately surrounding the erratics were deformed by the erratics when they were emplaced.

Overlying the lake sediments is a very stony till. The lower part of the till is brownish gray (5YR 5/1), owing to incorporation of some of the underlying lake sediments, but grades up into moderate yellowish brown (10YR 5/2); cobble-size till stones are most common. Most stones are poorly to moderately well rounded, but some are angular and subangular. In outcrop 1 the till is fairly compact. In outcrop 2 the matrix is very sandy, and the till is readily disaggregated. In outcrop 3 the till matrix is dominantly silt and clay resulting in an extremely hard, compact till.

Stratigraphic and structural relations of the lake sediments and till suggest that glacier ice overrode fine-grained sediments of an ice-margin lake, deformed them, and deposited till. It is not known whether this event marks a major ice advance in the southern Adirondacks or merely a local, relatively late readvance of the ice.

Folding in the glacial lake sediments is concentric, and beds show little or no change in thickness. Folds die out both upward and downward, and the lower structural discontinuity forms the décollement shown in Plate 1. The amplitude of the folds above the décollement is approximately 7 inches, and the wave length is 20 inches. Numerous normal, high-angle reverse, and thrust faults are present. In any one outcrop, different structures are complexly interrelated. Faults and folds are mutually deformed into intricate patterns. Sand fills the space between folds and the overthrust mass of lake sediments and shows evidence of having acted as an intrusive, indicating great mobility during deformation. Some folds have been refolded about other axes, and faults have been rotated about fold axes into positions very different from those in which they were formed. At first glance these outcrops suggest deformation in two or more separate events.

Two areas of several square feet each were chosen for study from outcrops 1 and 3, and within each area measurements were taken on all structural elements. All the structures were dealt with as encountered. The unconsolidated nature of the sediments permitted dissection of the structures in the outcrops systematically and viewing them in three dimensions.

Data on folds are presented on lower-hemisphere, equal-area projections in Figure 1a.

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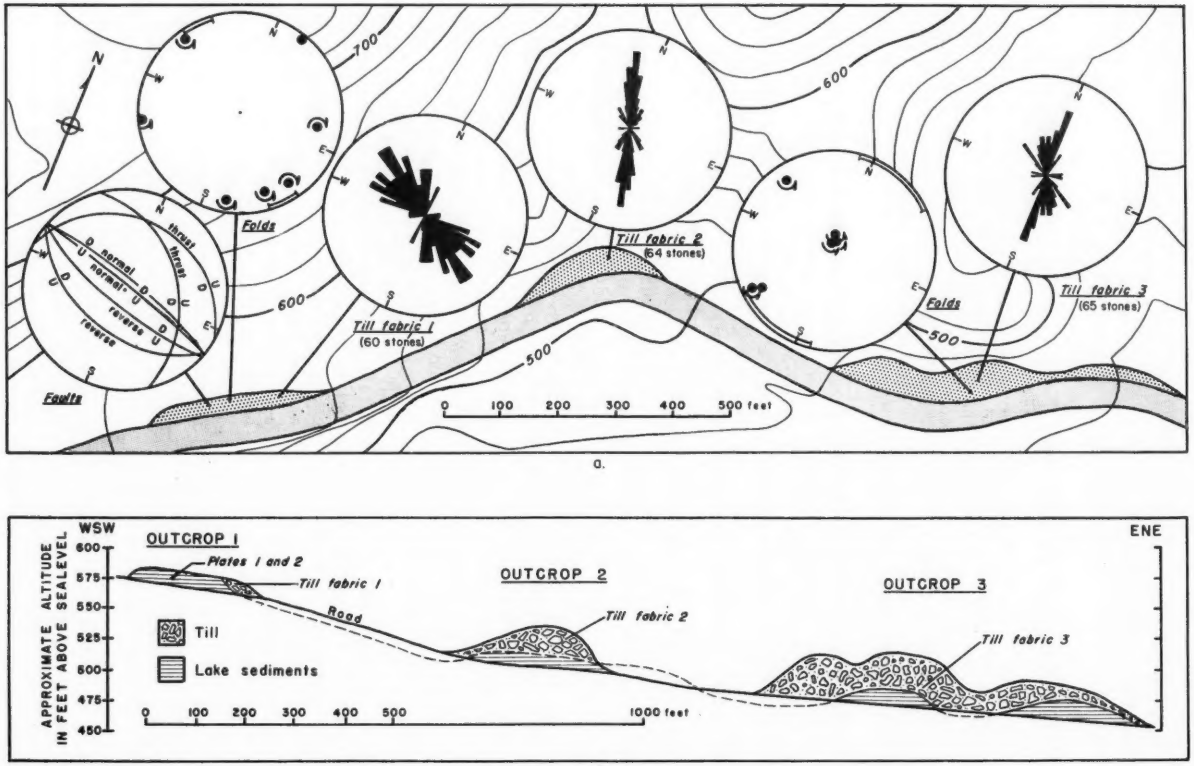

Figure 1. Outcrop area of glacial-lake sediments and overlying till

a. Plan view with till-fabric rose diagrams and lower-hemisphere, equal-area projections of fold axes and fault planes. Lines from diagrams point to corresponding outcrops. Topography is shown with 20-foot contour intervals.

b. Cross section of area

Folds from outcrop 1 have gently plunging axes, and the strikes of their axes differ greatly. Fold axes from outcrop 3 range in plunge from 5° to 80°, and their strikes are also quite divergent. This diversity in strike and plunge might be explained by superposed folding, in which case many more observations would be needed to separate structures of an earlier deformational event from those of a later event. However, these folds can all be related to a single deformation, if shear sense about the axes is considered. Most of the fold axes shown in Figure 1a are qualified by arcuate arrows. The arrows indicate shear in a rotational sense, clockwise and counterclockwise, about the fold axes. Where it was impossible to determine the shear sense about an axis, no sense is recorded.

Outcrop 1. Both shear senses are recorded—two are clockwise, and four are counterclockwise. In the southeastern quadrant of the diagram, the strikes of two fold axes with different shear senses are only 18° apart. Arcs have been added to the inner edge of the diagram to indicate this gap between strikes; they are shown in both the southeastern and northwestern quadrants. Across this 18-degree gap, the shear sense about folds is reversed. The four axes with counterclockwise shear lie on one side (SW) of the gap, and the two axes with clockwise shear lie on the opposite side (NE).

Outcrop 3. Fold axes show only one shear sense, counterclockwise, and their strikes range within 129°. This range is 51° short of covering the western hemisphere of the diagram. By using the fold-axis diagram from outcrop 1 as a model, it is expected that any axes found to fall within the eastern hemisphere would exhibit a clockwise shear. The folds shown allow a maximum arc of 51° to be drawn on the diagram; within this gap the shear sense about folds should reverse as in the fold-axis diagram from outcrop 1.

Comparison of the two diagrams shows that the 18-degree arc determined from folds in outcrop 1 lies 30° from the 51-degree arc of outcrop 3. If the two diagrams are combined, the disposition of fold axes and their shear directions makes it impossible to divide the diagram in half with axes of clockwise shear on one side and axes of counterclockwise shear on the other; instead, several shear reversals occur. These facts are interpreted to mean that the folds measured in outcrop 1 show a spatial orientation significantly different from the folds measured at outcrop 3.

Two thrust faults, two normal faults, and two high-angle reverse faults were measured in outcrop 1. Their spatial relations are presented on a lower-hemisphere, equal-area projection in Figure 1a.

The fabric of the till in each outcrop was measured, and the results are shown in rose diagrams in Figure 1a. Each fabric (numbered according to outcrop) shows a fairly strong unidirectional component, which is believed to parallel the direction of ice movement at the time of till deposition. Fabric 2 is the best developed of the three; fabrics 1 and 3 show a greater spread in the orientation of the long axes of till stones. In outcrop 2 the fabric was taken well above the basal contact of the till, whereas in outcrops 1 and 3 the fabrics were measured close to the boundary between till and lake sediments; a "bulldozing" effect during till emplacement may have caused the greater spread in orientations recorded in fabrics 1 and 3. The mean orientation in fabric 1 is about S. 55° E., in fabric 2 about S 20° E., and in fabric 3 about S. 10° E. The wide divergence of fabric 1 from the other two is believed to be due to the local influence of topography on ice movement; above outcrop 1 the general slope of the land is southeast, whereas near outcrops 2 and 3 the land slopes almost south. Striae elsewhere in the Glens Falls area indicate that the direction of ice movement was about S. 8°–15° E.

Comparison of the till-fabric and fold-axes diagrams of outcrop 1 shows that the mean orientation of stones in the till lies within the 18-degree arc determined from the folds. Similarly, diagrams of outcrop 3 show that the mean orientation of stones in the till lies within the 51-degree arc defined by the folds. However, stone orientation from the till at outcrop 1 does not lie within the arc determined from folds in outcrop 3, and, conversely, the mean orientation of stones from outcrop 3 does not lie within the arc determined from folds in outcrop 1. Furthermore, attempted correlation of till fabric 2 with either folds 1 or 3 is unsuccessful; in both cases the mean orientation of stones lies outside the arcs of 1 and 3. These correlations are interpreted as indicating that the folds were produced by movement of the glacier that deposited the overlying till. Although intuitively it would seem that fold axes should be either parallel or perpendicular to the direction of movement of the deforming force, they are oriented in various directions. The shear senses about the measured folds are compatible with overthrusting from

Figure 1. Outcrop area of glacial-lake sediments and overlying till
a. Plan view with till-fabric rose diagrams and lower-hemisphere, equal-area projections of fold axes and fault planes. Lines from diagrams point to corresponding outcrops. Topography is shown with 20-foot contour intervals.
b. Cross section of area

the north (fold-axis diagram from outcrop 3) and northwest (fold-axis diagram, outcrop 1) and underthrusting from the south and southeast (respectively). However, it is believed that overthrusting from the northwest and north caused the folding, because the ice probably moved southward and because the underlying beds are undisturbed.

The orientation of faults from outcrop 1 relative to till fabric 1 is difficult to analyze. Strikes of all faults are nearly parallel to the mean orientation of stones from till 1. A significant portion of the fault displacement may be strike-slip, but this was not verified at the outcrop.

The intricate interplay of faults and concentric folds above a décollement surface in the lake sediments is analogous to cover tectonics where the basement rocks remain undeformed below a folded and faulted sedimentary cover. Best known among large-scale structures of this type are those of the Jura Mountains, and the exposures of the lake sediments are believed to offer a model cross section through such structures (Pls. 1, 2). Significant points of difference are, of course, that the mechanism that deformed the glacial lake sediments was overthrusting by glacial ice and not gravity sliding, basement control, or lateral compression (from the Alps), the principal hypotheses for the Jura.

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Figure 1. Deformed glacial lake sediments above sediments containing ice-rafted stones and below undeformed sediments of the thrust sheet. Strike of face is N.46°E.

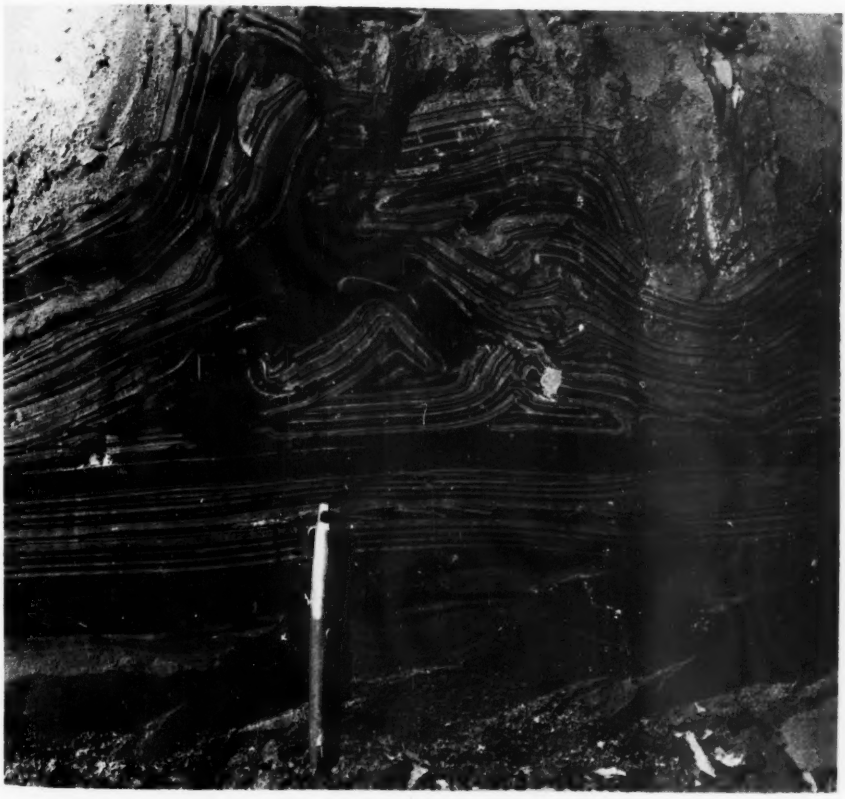

Figure 2. Detail of structural element from Figure 1

GLACIAL-LAKE SEDIMENTS FROM OUTCROP 1

HANSEN *ET AL.*, PLATE 1
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Figure 1

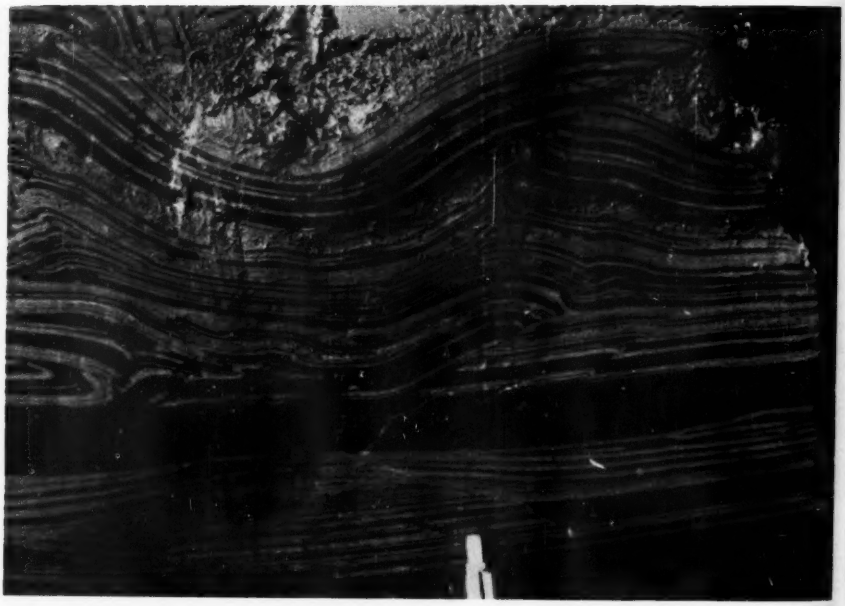

Figure 2

DETAILS OF STRUCTURAL ELEMENTS FROM FIGURE 1 OF PLATE 1

BARRY VOIGHT

CONSOLIDATION OF SEDIMENTS

Abstract: Edwin L. Hamilton suggests that geologists should adopt the terminology of soil mechanics in describing consolidation of sediments. G. H. Davis takes issue with this statement. This note discusses and corrects Davis' definition of consolidation as it applies to soil mechanics.

Davis (1960, p. 1728) has defined "consolidation," as used in soil mechanics, as the process by which earth materials become more dense through application of load. The writer believes this definition is not consistent with the present viewpoint of soil engineers.

The Glossary of Geology and Related Sciences (American Geological Institute, 1957) defines the term as follows:

"Consolidation: 1. In Geology, any or all of the processes whereby loose, soft, or liquid earth materials become firm and coherent. (Stokes—Varnes). 2. In soil mechanics it refers to the adjustment of a saturated soil in response to increased load, and involves the squeezing of water from the pores and decrease in void ratio.[1] (Stokes—Varnes)"

In part 2 of the definition, "consolidation" involves the squeezing of pore water and hence is limited by basic assumption to saturated soil (Hough, 1960, p. 102). Davis did not include this limitation to saturated soil; thus, part 2 differs from his concept of the term as used in soil mechanics. He used part 1 of the above definition to define correctly the term as used in geology.

Taylor (1948, p. 212) writes:

"The gradual process which involves, simultaneously, a slow escape of water and a gradual compression . . . is called consolidation. This definition is a general one, valid for the three-dimensional as well as the one-dimensional case. There has been some controversy regarding the proper definition of consolidation as the term is understood in soil mechanics. It is suggested that the definition given herein, according to which consolidation is a gradual process involving drainage, compression, and stress transfer, is preferable to the definition used by some soil engineers, according to which consolidation is merely compression under static load."

[1] The void ratio is the ratio of the volume of the voids to the volume of the solid substance (Terzaghi and Peck, 1948, p. 26).

Three factors that contribute to a decrease in volume for a soil under compressive stress are listed (Taylor, 1948, p. 209): 1. A compression of the solid matter; 2. A compression of liquid and gas within the voids; 3. An escape of liquid and gas from the voids.

The components of total compression due to a decrease in volume of the solid matter and of the liquid (*i.e.*, pore water) may be considered negligible, as both are relatively incompressible. Thus if a soil is saturated, compression is realized entirely by the escape of water from the pores, which permits repacking of the soil grains. To the soil engineer, this process is "consolidation" (Tschebotarioff, 1951, p. 95). In a nonsaturated condition the soil may compress as a result of an escape of gas (*i.e.*, air) from the voids and/or a compression of gas within the voids. As Taylor (1948, p. 210) remarks, this condition is far more complex, since a small amount of gas within the pores may permit appreciable compression. In this instance the compression occurs even though there is no escape of pore water, and hence, to the soil engineer, this process is not "consolidation."

It is not within the scope of this discussion to attempt a complete explanation of the concept of "consolidation." A better understanding may be obtained from any one of a number of excellent references on soil mechanics listed in the Appendix.

Hamilton (1959, p. 1402) states that geologists should not use "consolidated sediment" when they mean "lithified sediment." He implies that the definition used by soil engineers should be used by geologists, in preference to the established geological definition. The writer believes, in agreement with Davis (1960, p. 1728), that established geological definitions should not be readily abandoned in favor of the specialized definitions employed in other fields. The reasons presented by Davis

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are clear and meaningful. However, should the geologist's work involve these other fields, the possibility of mistaken interpretation due to terminology must be considered. Hamilton did consider this point in his paper, in which the interdisciplinary terminology is very adequately defined.

Acknowledgments

This note was read in manuscript by Dr. Raymond C. Gutschick, Mr. William M. Fairley, and Mr. Leroy D. Graves. The writer is indebted to them for their helpful suggestions.

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Terzaghi, K. and Peck, R. B., 1948, Soil mechanics in engineering practice: New York, John Wiley and Sons, 566 p.

Tschebotarioff, G. P., 1951, Soil mechanics, foundations, and earth structures: New York, McGraw-Hill, 655 p.

Appendix:

The following texts are suggested as useful references for the basic principles of soil mechanics:

Hough, B. K., 1957, Basic soils engineering: New York, Ronald Press, 513 p.

Krynine, D. P., 1947, Soil mechanics, 2d ed.: New York, McGraw-Hill, 511 p.

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RECENT CHEMICAL ANALYSES OF WATERS FROM SEVERAL CLOSED-BASIN LAKES AND THEIR TRIBUTARIES IN THE WESTERN UNITED STATES

Abstract: Some of the classic closed-basin lakes of the western United States have been resampled, and the waters have been analyzed by modern wet-chemical methods. Included are waters from Borax and Little Borax lakes and Mono Lake in California; Big Soda, Pyramid, and Walker Lakes in Nevada; Abert Lake, Oregon; and Great Salt Lake, Utah. Tributary streams and springs have also been sampled and are reported upon.

Many discussions of closed-basin lakes have appeared in the literature. Clarke (1924, p. 156-180) included a chapter on them in his *Data of geochemistry*. Rankama and Sahama (1950) and Hutchinson (1957) also considered the chemistry of these complex waters. Hutchinson (1937) studied several lakes in the western Lahontan Basin and discussed theoretical aspects of the chemistry of the waters. He also showed the effect of return flow from irrigation upon the chemical composition of the lake waters.

The present note gives the results of recent chemical analyses of waters from several closed-basin lakes in the western United States and analyses of some of the stream and spring waters tributary to them. The locations of the lakes are shown in Figure 1, and analytical data are given in Table 1.

The samples were collected and analyzed by the Water Resources Division of the U.S. Geological Survey as part of a geochemical study of natural waters in various environments and in relation to weathering. The waters were analyzed by methods current in the Geological Survey (Rainwater and Thatcher, 1960), which incorporate many advances in analytical instrumentation and reagents in the 50 or 60 years since the collection of Clarke's (1924) data. Most of the lake waters reported on by Clarke were collected and analyzed in the late nineteenth century.

The evolution of the composition of natural waters is complex and is influenced by many factors. White (1957a; 1957b) contributed much information about the composition of waters from springs and thermal springs; Hem (1959) discussed in general terms many of the factors that cause changes in the chemical composition of waters; and Baas-Becking, Kaplan, and Moore (1960) considered the biological effects of natural environments on pH-Eh conditions.

Principal factors affecting the composition of closed-basin waters are interrelations of temperature, precipitation, and evaporation, including long-term effects of concentration by evaporation in some lakes, such as Great Salt Lake, and the amount and composition of influent waters from streams and springs; also involved are the effects of the biota, the basin sediments, and the kinds of rock that compose the drainage basin. Little is known of these variables, but some of the factors that have affected the composition of the waters are noted briefly in the following sketches of the closed-basin environments in the Western United States.

CALIFORNIA: The waters from closed-basin lakes of California (Table 1, nos. 117, 344, 393, 455) are generally similar in their chemical composition, although there is a wide range in concentrations. The waters are alkaline, the dissolved solids are fairly high, and calcium and magnesium contents are low; magnesium exceeds calcium in all waters sampled in this group.

The Borax Lakes (nos. 117, 344, 393) appear as chlorocarbonate types. The differences in concentration between the two samples from Little Borax Lake may result from dilution by increased runoff and movement of ground water during the winter and early spring. The diluting waters appear to be richer in calcium, bicarbonate, and sulfate, yet notably less concentrated in silica. The mechanisms by which

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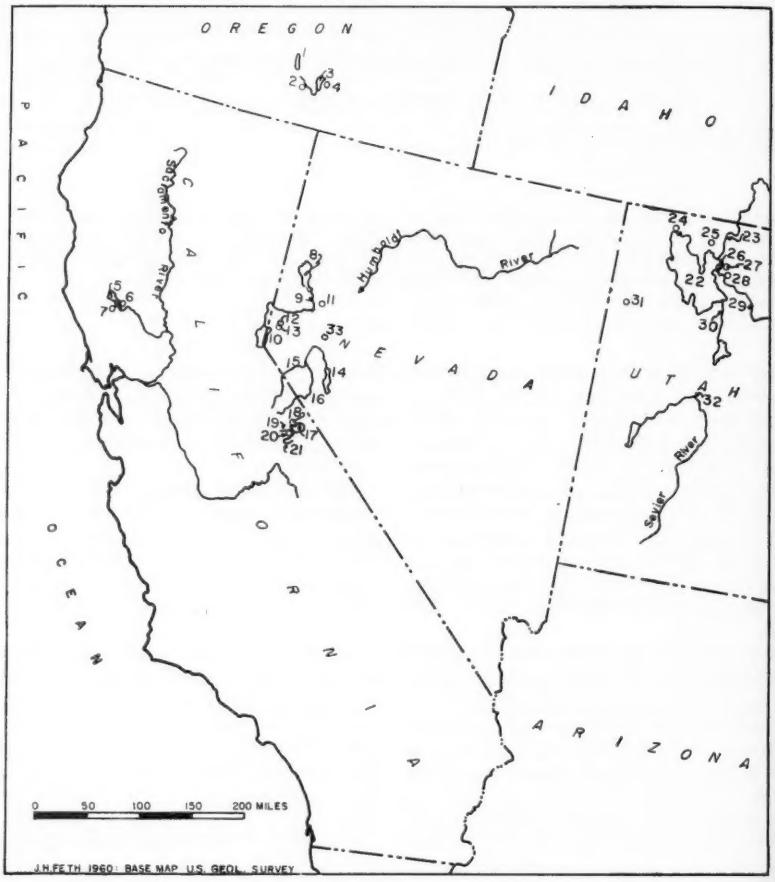

Figure 1. Map of western United States showing places referred to in the text

CALIFORNIA:
- Borax Lake (6)
- Clear Lake (5)
- Leevining Creek; Rhinedollar Reservoir, and Saddlebag Lake are in this drainage system (19)
- Little Borax Lake (7)
- Mono Lake (17)
- Mono Lake, cold spring on West shore (20)
- Mono Lake, hot spring on north shore (18)
- Rush Creek; June Lake is in this drainage system (21)

NEVADA:
- Big Soda Lake (33)
- Big Washoe Lake (13)
- Closed-basin sump northeast of Fernley (11)
- East Walker River (16)
- Lake Tahoe (10)
- Little Washoe Lake (12)
- Pyramid Lake (8)

NEVADA: (*Continued*)
- Truckee River between Wadsworth and Nixon (9)
- Walker Lake (14)
- West Walker River (15)

OREGON:
- Abert Lake (3)
- Abert Lake; spring on east side (4)
- Chewaucan River (2)
- Summer Lake (1)

UTAH:
- Bear River (23)
- Bonneville Salt Flats; brine sample (31)
- El Monte Hot Spring (28)
- Great Salt Lake (22)
- Hooper Hot Spring (26)
- Jordan River (30)
- Locomotive Spring (24)
- Sevier River near Lyndyll (32)
- Stinking Spring west of Corinne (25)
- Utah Hot Springs (27)
- Weber River (29)

the water changes from the concentration observed in the spring to that in the summer have not been determined.

Waters from Borax Lake and Little Borax Lake are the product of a series of physical and chemical processes largely unidentified. Apparently there is an abundance of carbonate so great that precipitation of calcium carbonate has occurred without significantly depleting the total bicarbonate and carbonate content. A neighboring body of fresh water, Clear Lake (U.S. Geological Survey, 1959, p. 300–304), which is not in a closed basin, contains calcium and bicarbonate as the main constituents, along with lesser amounts of sodium and potassium. Chloride and sulfate are less abundant than bicarbonate and are present in about equal quantities in Clear Lake.

Big Borax Lake lies on the east side of Clear Lake (Fig. 1) and is surrounded larely by metasedimentary rocks of the Franciscan formation. Little Borax Lake, on the south-west side of Clear Lake, lies in Quaternary volcanic rocks. Some differences in mineralization of waters from the three lakes may be due to differences in geologic environment, but no correlation has been seen. Clear Lake discharges through Cache Creek to the Sacramento River, whereas the Borax Lakes occupy basins that have no surface outflow; this is the most obvious difference.

Several springs whose waters are of varied and largely unknown composition discharge into and around Mono Lake (no. 455) and into its tributaries (Feth, 1959). Water from Mono Lake is dominated by sodium among the cations and has noteworthy amounts of all the major anions—chloride, sulfate, carbonate, and bicarbonate. Several of the tributaries and springs that feed Mono Lake discharge water of a comparable chemical character (Table 1, nos. 446, 447, 449, 450, 452, 451, 453, 454, 456, 21261).

The geologic environment also is complex, including granite, Quaternary volcanic rock, and Pleistocene sedimentary rocks. No attempt is made here to evaluate either the effects of the geology or the influent waters.

NEVADA: The waters of closed-basin lakes in Nevada (nos. 1092, 1157, 1090, 804, 921, 935, 936, 1089) show some similarity in chemical composition to those in California. The exceptions in the group considered here are Big and Little Washoe lakes, which are not true closed-basin lakes. The land surrounding them is in part cultivated, and in wetter periods there is drainage out of the basin. The relatively low mineralization, the low pH, the excess of calcium over magnesium, and the relatively high concentration of silica are not typical of waters from closed-basin lakes. The two samples of water from Big Washoe Lake (nos. 1092, 1157) represent, respectively, spring and early winter and reflect dilution by influent water. Both Big and Little Washoe lakes receive considerable amounts of influent water. Their present compositions are cited as examples of water from intermittently closed basins because these compositions may result from loss of salts spilling from closed-basin lakes during extremely wet periods.

Pyramid, Walker, and Big Soda lakes lie in areas of volcanic rocks and weakly consolidated or unconsolidated Pleistocene sediments and are within the area once covered by ancient Lake Lahontan. Waters from all these lakes are similar in chemical composition, showing notable variation only among the anions—sulfate in particular. The similarity of chemical composition is particularly striking, as all these lakes receive various amounts of return flow from irrigation. Big Soda Lake is fed entirely by ground water, whereas Pyramid and Walker lakes are fed by major streams. The chemical composition of water from the Truckee River, flowing from Lake Tahoe into Pyramid Lake, and that of the Walker River, which supplies Walker Lake, are shown in Table 1 (nos. 944, 945, 466, 467, 805).

The water of a closed-basin sump (no. 1089) just north of Fernley, in a comparable geologic environment, is twice as saline as Big Soda Lake, yet has a significantly lower pH, is very low in bicarbonate content, and contains appreciable quantities of calcium and magnesium.

Some return flow from irrigation reaches the several closed basins named, but, except in Big Soda Lake, it is probably not a large percentage of the total inflow. The principal present-day contributor to mineralization of the waters tributary to all these lakes is probably the unconsolidated material making up the Lahontan Valley group. This influence is visible, for instance, in the increased concentration of water of the Truckee River between Wadsworth and Nixon, Nevada (nos. 944, 945), a reach in which there is no irrigated land but where the river channel is cut into, and underlain by, sediments of the Lahontan Valley group. The increase in total concentration in the 15-mile stretch from Wadsworth to Nixon is striking and is probably the result of discharge of highly concentrated ground water into the river. The influent waters probably acquire their mineral content

largely from the more soluble portions of the unconsolidated sediments of the Lahontan Valley group. Further field work would be required, however, to determine the actual source of the mineralization.

The water level in Big Soda Lake has risen about 15 feet since inception of irrigation in the surrounding areas, according to records of the U.S. Bureau of Reclamation (D. A. Barker, oral communication, 1960), reflecting a comparable rise of the water surfaces in wells in the irrigated areas. The water sample collected from the lake in 1958 had only about one-fifth the concentration of that reported by Clarke (1924).

OREGON: Abert Lake is shallow and lies in the eastern part of the Chewaucan Basin of south-central Oregon (Fig. 1). It is fed by the Chewaucan River and several springs (Table 1, no. 725). The predominant rocks of the lake basin "* * * consist of basalt and tuffs or the disintegration and erosion products of these * * *" (Van Winkle, 1914, p. 117). Baldwin (1959, p. 100) reported Miocene to Recent basaltic flows, pyroclastic rocks, and alluvial sediments in the basin. The Chewaucan Basin is of interest to students of closed-basin lakes, as Abert Lake and Summer Lake to the northwest lie in the depression of a larger basin that once during Quaternary time contained Lake Chewaucan.

There are volcanic rocks in the Lahontan Basin also, but their influence on the chemical composition of waters in the basin is probably not so pronounced as that suggested by the much higher silica content of Abert Lake. The apparent effect of runoff in diluting the water of Abert Lake may be seen by comparing the sample collected in October (no. 533) with that collected the following May (no. 724). Much of the runoff presumably originates in the Chewaucan Marsh. Water derived from precipitation and the discharge of springs is delayed in transit, concentrated by evaporation, and altered in chemical character during prolonged contact with swamp muck and vegetation.

Abert Lake appears to be a chlorocarbonate type of closed-basin lake, comparable to the Borax Lakes of California. Again, there apparently has been sufficient carbonate in the water to precipitate the calcium and magnesium without reducing the bicarbonate and carbonate content appreciably.

UTAH: Great Salt Lake, a remnant of ancient Lake Bonneville, has received attention as a hydrologic phenomenon and a challenging problem in water chemistry. The early survey of Gilbert (1890) remains the classic description of the basin, the lake, and its shore features. More recent papers have discussed the hydrology (Eardley, 1938; Eardley and others, 1957). The geology of the strata deposited in Lake Bonneville has been described by many. The bibliography presented by Hunt and his associates (1953, p. 92–96) lists many of the pertinent references.

The chemical composition of Great Salt Lake (no. 691) fluctuates considerably from year to year and season to season. It varies, also, with depth and location of sampling (Connor and others, 1958, p. 256). There is a persistent similarity, however, in the chemical character of the water throughout the lake. Great Salt Lake water contains primarily sodium chloride which has accumulated from several sources (Feth, 1959). Apparently it has reached a concentration that will not permit many other constituents in solution in large concentrations. In the colder parts of some years, crystals of sodium sulfate (mirabilite) form in and around the edge of the lake. These redissolve in summer when the water temperature rises (Eardley, 1938, p. 1322–1323).

Analyses of waters tributary to Great Salt Lake also are shown in Table 1 (nos. 689, 690, 961, W-78, X-42, 1186, 1187, 1188, W-54). Most of the springs are rich in sodium and chloride, adding to the vast amounts already in the lake. A variety of mineral constituents is added through the Bear, Weber, and Jordan rivers. The Sevier River is included in this group becuase it drains an area that was tributary to Lake Bonneville at various stages of that pluvial lake. The high content of dissolved solids carried by the Sevier (U.S. Geol. Survey, 1959, p. 199–201) is picked up mostly from bedded halite and gypsum underlying many square miles of the Sevier River valley.

Similar waters occur throughout the Great Salt Lake Basin. A typical brine is sample 692 collected on the Bonneville salt flats. It is even more concentrated than water from Great Salt Lake but is chemically similar. The much increased calcium content may correspond to the decrease in bicarbonate.

The Bonneville Basin is characterized by a diversity of geologic terranes. The deposits of salt and gypsum have been mentioned. Volcanic rocks are prominent in the northern and western parts of the basin, and sedimentary rocks of Paleozoic, Mesozoic, and Cenozoic age are abundant throughout the area. Igneous and metamorphic rocks underlie large areas in the eastern part of the basin. Thus, geologic control of the mineralization of the water is complicated.

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No.

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1089

533[2 3 6]
724[2 3 6]
725

691[2 3 7]
689
690
961
W-78[8]
X-42[8]
1186[2 7]
1187[2 7]
1188[2 7]
W-54[8 2 7]
692

No.	Sample	Date of collection (Discharge)	Temperature, °F	Silica (SiO_2)	Aluminum (Al)	Iron (Fe)	
117	Little Borax Lake, Lake County	10–6–56	Cold	20	..	..	
344	Little Borax Lake, Lake County	3–28–57	..	.9	..	..	
393	Big Borax Lake, Lake County	3–26–57	..	4.3	1.1	0.00	0
455[2 3]	Mono Lake[4], Mono County	9–20–57	62	14	..	.60	
446	Rhinedollar Reservoir, Mono County	9–19–57	..	..	..	..	
447	Saddlebag Lake, Mono County	9–19–57	..	1.8	..	.00	
449	Warren Fork of Leevining Creek, Mono County	(200–250 gpm) 9–18–57	48	..	..	..	
450	Leevining Creek, 1 mile from U.S. 395, Mono County	(500 gpm) 9–18–57	50	..	..	..	
452	June Lake, Mono County	9–19–57	64	..	..	..	
451	Rush Creek, head of Grant Lake, Mono County	(10–15 cfs) 9–19–57	61	..	..	..	
453	Rush Creek, 4 miles N. of Leevining, Mono County	(2 gpm) 9–19–57	59	..	..	..	
454	Spring, west end of Mono Lake, Mono County	(2 gpm) 9–20–57	53	..	..	..	
456	Hot Spring, Black Point, Mono Lake, Mono County	(100 gpm) 9–20–57	144	73	..	.00	
21261	Spring[5] southwest corner Mono Lake, Mono County	(1 gpm) 11–15–56	53	37	..	..	
1092	Big Washoe Lake, Washoe County	4–29–59	..	46	.00	.00	
1157	Big Washoe Lake, Washoe County	11–10–59	..	61	..	..	
1090	Little Washoe Lake, Washoe County	4–24–59	63	34	.00	.00	
804[2 3]	Big Soda Lake, Churchill County	8–28–58	..	3.3	.89	.10	
921[2 3]	Pyramid Lake, Washoe County	10–22–58	62	1.3	.68	.00	
944	Truckee River at Wadsworth	11–21–58	..	23	.00	.00	
945	Truckee River at Nixon	11–21–58	..	28	..	..	
935[2 3]	Walker Lake, north end, Mineral County	11–20–58	..	5.6	1.5	.00	
936[2 3]	Walker Lake, south end, Mineral County	11–20–58	..	3.4	1.5	.00	
467	East Walker River, above Nordyke	9–21–57	62	..	..	..	
466	West Walker River, below Smith, Lyon County	9–21–57	62	..	..	..	
805	West Walker River, north of Coleville	(3–5 cfs) 8–29–58		19	.00	.00	
1089	Closed-basin sump, Lyon County	4–23–59	..	9.1	2.0	.00	
533[2 3 6]	Abert Lake, Lake County	10–29–57	44	140	.68	.00	
724[2 3 6]	Abert Lake, Lake County	5–18–58	65	69	.23	.00	
725	Spring, east shore Abert Lake, Lake County	5–18–58	..	36	.22	.11	
691[2 3 7]	Great Salt Lake, Salt Lake County	4–6–58	48	6.2	2.4	.00	
689	Stinking Hot Springs, Box Elder County	4–5–58	118	48	1.0	.03	
690	Utah Hot Springs, Box Elder County	5–4–58	134	38	.46	.42	1.9
961	Locomotive Spring, Box Elder County	11–20–58	61	36	.17	.06	
W-78[8]	Hooper Hot Springs, Weber County	9–15–53	140	35	..	..	
X-42[8]	El Monte Hot Spring, Weber County	4–27–43	135	..	..	..	
1186[2 7]	Weber River, Plain City	1–6–60	37	..	..	..	
1187[2 7]	Bear River, Corrinne	1–6–60	35	..	..	..	
1188[2 7]	Jordan River, Salt Lake City	1–5–60	50	..	..	..	
W-54[8 2 7]	Weber River, 24th St. Viaduct, Ogden	8–13–54	..	12	..	..	
692	Brine beside U.S. Highway 40, Bonneville salt flats 10± miles east of Wendover	4–6–58	..	7.4	1.8	.00	

[1] Turbidimetric sulfate

[2] *See* Clarke (1924, Chapter V)

Silica (SiO_2)	Aluminum (Al)	Iron (Fe)	Manganese (Mn)	Copper (Cu)	Lead (Pb)	Zinc (Zn)	Arsenic (As)
20	..	..	..	..	..	..	..
.9	..	..	..	..	..	..	..
4.3	1.1	0.00	0.00	0.02	0.00	0.20	0.01
14	..	.60	..	..	..	..	..
..	..	..	..	..	..	..	..
1.8	..	.00	..	..	..	..	..
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
73	..	.00	..	..	..	..	..
37	..	..	..	..	..	..	..
46	.00	.00	.00	..	..	..	..
61	..	..	..	..	..	..	..
34	.00	.00	.00	..	..	..	..
3.3	.89	.10	.00	..	..	..	.00
1.3	.68	.00	.00	..	..	..	..
23	.00	.00	..	..	..	..	..
28	..	..	..	..	..	..	..
5.6	1.5	.00	.00	..	..	..	..
3.4	1.5	.00	.00	..	..	..	.00
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
19	.00	.00	.00	..	..	..	.00
9.1	2.0	.00	.01	..	..	..	..
40	.68	.00	.09	.00	.00	.00	.56
69	.23	.00	..	..	..	..	..
36	.22	.11	.00	..	..	..	.00
6.2	2.4	.00	.00	..	..	..	.07
48	1.0	.03	.00	.00	.01	.00	.00
38	.46	.42	1.9	.00	.01	.00	.00
36	.17	.06	.00	.00	.00	.01	.00
35	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
..	..	..	..	..	..	..	..
12	..	..	..	..	..	..	..
7.4	1.8	.00	.00	..	..	..	.00

c sulfate
924, Chapter V)

[3] *See* Gilbert (1890, p. 176)
[4] Values reported are mg/l

TABLE 1. CHEMICAL ANALYSES OF WATERS FROM SELECTED CLOSED-BASIN LAKES AND THEIR TRIBUTARIES IN WESTERN

Constituents and properties reported as parts per million, unless otherwise noted

Arsenic (As)	Calcium (Ca)	Magnesium (Mg)	Strontium (Sr)	Sodium (Na)	Potassium (K)	Lithium (Li)	Ammonia nitrogen as NH_4	Bicarbonate (HCO_3)	Carbonate (CO_3)	Sulfate (SO_4)
					CALIFORNIA					
..	8.0	24	5.3	3390	731	1.2	..	626	3770	1
..	6.4	8.7	..	1330	285	.3	..	970	1220	
0.01	4.0	25	..	6140	295	.8	0.5	1160	2650	2
..	4.3	37	..	21,700	1150	..	.9	5230	11,200	787
..	..	..	..	.7	.0	..	..	..	..	
..	2.4	.0	..	.8	.5	..	.0	7	0	
..	..	..	..	1.0	.4	..	..	..	..	
..	..	..	..	2.5	.5	..	..	..	..	
..	..	..	..	27	4.5	..	..	..	..	
..	..	..	..	1.1	.4	..	..	..	..	
..	..	..	..	12	2.5	..	..	..	..	
..	..	..	..	100	9.0	..	..	..	..	
..	14	4.6	..	450	6.7	..	1.9	430	0	9
..	33	4.1	..	14	3.4	..	..	148	0	
					NEVADA					
..	38	14	..	62	9.6	.05	.0	325	0	1
..	41	18	..	114	16	..	..	464	10	3
..	28	7.0	..	30	5.0	.04	.0	189	0	
.00	7.9	195	.5	8610	39	.00	.0	1250	1360	622
..	7.2	110	.7	1610	99	.97	.3	980	186	28
..	16	1.9	..	20	..	..	.0	103	0	1
..	33	13	..	43	3.5	..	..	119	0	6
..	7.2	105	..	2330	95	.00	.0	1360	419	150
.00	6.8	105	..	2330	96	.00	.0	1370	410	157
..	..	..	..	25	4.8	..	..	..	..	
..	..	..	..	33	3.4	..	..	..	..	
.00	16	4.6	..	10	2.2	.00	.0	88	0	
..	463	751	..	19,300	510	.07	.3	382	0	10,00
					OREGON					
.56	1.2	1.5	.0	8370	295	.00	.5	2160	3230	39
..	4.2	12	..	3800	176	.00	.3	1470	1420	26
.00	9.6	7.0	..	14	2.2	.00	.0	64	0	
					UTAH					
.07	241	7200	15±5	83,600	4070	23	.0	251	0	16,40
.00	946	297	..	12,600	571	6.9	40	324	0	11
.00	1140	70	..	7030	904	9.9	5.4	192	0	18
.00	127	73	4.5	1030	48	.5	.01	196	0	11
..	536	92	..	2520	285	..	..	234	0	3
..	356	11	..	3035[5]	..	..	..	203	0	10
..	71	25	..	..	..	..	..	352	0	
..	13	86	..	..	..	..	..	428	0	
..	168	66	..	..	..	..	..	362	0	
..	62	14	..	15	2.0	..	..	222	8.7	2
.00	1770	1360	..	96,200	2930	18	.3	40	0	377

90, p. 176)
d are mg/l

[5] Analysis by U. S. Geol. Survey, Sacramento, Calif., Qual
Laboratory

SELECTED CLOSED-BASIN LAKES AND THEIR TRIBUTARIES IN WESTERN UNITED STATES
erties reported as parts per million, unless otherwise noted

	Potassium (K)	Lithium (Li)	Ammonia nitrogen as NH_4	Bicarbonate (HCO_3)	Carbonate (CO_3)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F)	Bromide (Br)	Iodide (I)	Nitrite (NO_2)
		CALIFORNIA									
	731	1.2	..	626	3770	10[1]	905	..	..	..	..
	285	.3	..	970	1220	5.8	365	..	..	..	..
	295	.8	0.5	1160	2650	20	5760	1.0	19	8	0.00
	1150	..	.9	5230	11,200	7870	14,500	44	35	6	.04
.7	.0	..	..	..	..	..	.5	..	..	..	..
.8	.5	..	.0	7	0	1.8	.2	.0	.0	.0	.00
.0	.4	..	..	..	..	..	.5	..	..	..	..
.5	.5	..	..	..	..	..	.9	..	..	..	..
	4.5	..	..	..	..	..	4.2	..	..	..	..
.1	.4	..	..	..	..	..	.9	..	..	..	..
	2.5	..	..	..	..	..	1.8	..	..	..	..
	9.0	..	..	..	..	..	20	..	..	..	..
	6.7	..	1.9	430	0	92	324	13	2.1	.4	.00
	3.4	..	..	148	0	8.6	3.2	..	..	..	..
		NEVADA									
	9.6	.05	.0	325	0	19	7.8	1.4	..	.02	.00
	16	..	..	464	10	37	15	2.1	..	..	..
	5.0	.04	.0	189	0	8.0	3.6	.7	..	.02	.00
	39	.00	.0	1250	1360	6220	7570	7.9	18	1.1	.00
	99	.97	.3	980	186	280	1920	1.5	5.4	.6	.00
	..	..	.0	103	0	14	5.5	..	..	..	..
	3.5	..	..	119	0	64	51	..	..	..	..
	95	.00	.0	1360	419	1500	1630	15	3.6	.4	.00
	96	.00	.0	1370	410	1570	1640	15	3.3	.6	.02
	4.8	..	..	..	..	..	5.9	..	..	..	..
	3.4	..	..	..	..	..	13	..	..	..	..
	2.2	.00	.0	88	0	6.0	6.0	.3	.0	.2	.01
	510	.07	.3	382	0	10,000	25,200	4.8	32	1.3	.03
		OREGON									
	295	.00	.5	2160	3230	397	7440	13	35	2.2	.05
	176	.00	.3	1470	1420	265	3520	1.3	..	..	.00
	2.2	.00	.0	64	0	9.2	17	.1	.0	.0	.00
		UTAH									
	4070	23	.0	251	0	16,400	140,000	4.2	86	1.4	.01
	571	6.9	40	324	0	111	21,600	1.9	15	1.3	.00
	904	9.9	5.4	192	0	189	13,300	3.2	8.2	.2	.00
	48	.5	.01	196	0	119	1910	.3	..	.008	.00
	285	..	..	234	0	36	4370	.7	..	..	..
9	..	..	..	203	0	103	5050	3.0	..	..	..
..	..	..	..	352	0	..	85	..	.0	.02	.00
..	..	..	..	428	0	..	302	..	.2	.02	.00
..	..	..	..	362	0	..	269	..	.2	.04	.00
	2.0	..	..	222	8.7	29	18	..	..	..	..
	2930	18	.3	40	0	3770	156,000	1.3	35	.9	.01

[5] Analysis by U. S. Geol. Survey, Sacramento, Calif., Quality of Water Laboratory

[6] *See* Van Winkle (1914, p. 120)

[7] *See* Connor, Mitchell, and others (1

Nitrite (NO_2)	Nitrate (NO_3)	Orthophosphate (PO_4)	Boron (B)	Dissolved solids: Calculated	Dissolved solids: Residue on evap. at 180°C	Hardness as $CaCO_3$	Noncarbonate hardness as $CaCO_3$	Specific conductance (micromhos at 25°C)	pH	Density at 20°C
..	..	..	360	9540	..	120	0	12,400	10.0	1.006
..	0.8	..	150	3850	..	52	0	5780	9.6	..
0.00	13	6.6	290	15,800	16,200	112	0	22,100	9.7	1.010
.04	16	..	350	59,500	61,400	163	0	60,300	9.7	..
..	..	..	..	..	..	..	..	18	7.3	..
.00	.1	..	..	..	..	6	0	17	6.1	..
..	..	..	..	..	..	..	..	64	7.6	..
..	..	..	..	..	..	..	..	67	7.7	..
..	..	..	..	..	..	..	..	228	8.5	..
..	..	..	..	..	..	..	..	24	7.1	..
..	..	..	..	..	..	..	..	162	7.4	..
..	..	..	..	..	..	..	..	580	8.4	..
.00	2.0	..	7.4	1200	1190	54	0	1960	8.0	..
..	1.3	..	.09	178	..	100	0	253	..	..
.00	1.8	1.1	.21	361	387	150	0	546	7.8	..
..	1.2	..	..	546	..	184	0	804	8.3	..
.00	1.2	.60	.19	211	224	100	0	318	7.5	..
.00	2.2	12	51	24,700	24,200	164	0	31,800	9.6	1.017
.00	.0	.06	14	4700	..	470	0	7930	8.9	..
..	.5	..	.43	132	..	67	0	236	7.5	..
..	.5	..	..	295	..	134	39	505	7.5	..
.00	.2	1.4	13	7300	7030	448	0	10,300	9.2	1.003
.02	.1	1.5	15	7040	7040	449	0	10,400	9.3	1.003
..	..	..	..	..	..	..	..	316	7.8	..
..	..	..	..	..	..	..	..	359	8.1	..
.01	.1	.04	.25	105	..	..	..	168	7.5	..
.03	2.1	.7	51	56,500	56,500	4250	3930	62,100	7.6	1.036
.05	3.8	33	32	21,100	21,600	4	0	30,500	9.8	1.016
.00	2.6	12	15	10,000	10,400	60	0	15,900	9.8	1.005
.00	1.6	.32	.10	129	..	53	0	175	6.9	..
.01	.0	.50	30	254,000	262,000	35,700	30,000	168,000	7.4	1.197
.00	.0	.00	3.6	36,500	37,300	3560	3320	53,900	6.7	1.025
.00	.0	.00	5.1	22,800	23,900	3030	2980	34,000	7.3	1.013
.00	1.6	.01	.39	3440	..	618	452	6120	7.6	..
..	2.0	..	1.2	..	9310	1710	1520	14,900	7.1	..
..	..	..	..	..	8650	931	767	..	..	..
.00	..	..	..	..	..	280	0	840	7.4	..
.00	..	..	..	..	..	384	36	1650	7.4	..
.00	..	..	..	..	..	690	394	2090	6.9	..
..	1.3	..	.00	..	265	212	16	463	8.5	..
.01	4.8	.00	4.2	262,000	271,000	10,000	..	180,000	7.1	1.201

, p. 120)
and others (1958, p. 250–272)

[8] Analyses by U. S. Bureau of Reclamation, Salt Lake City, Utah
[9] Potassium included in sodium value

D. J. CEDERSTROM

ORIGIN OF A SALT-WATER LENS IN PERMAFROST AT KOTZEBUE, ALASKA

Abstract: Frozen sediments were found to a depth of 238 feet in the drilling of a 325-foot well at Kotzebue, Alaska. Between 79 and 86 feet, however, highly saline water was found in a gravel lens. The writer suggests that the salt water originated by fractionation by freezing. Analyses of this water and of slightly saline water from below the permafrost are given.

In the summers of 1949 and 1950 a test well was drilled at Kotzebue, Alaska, in an endeavor to locate a source of potable water that might be utilized by the 500 townspeople (largely Eskimos) and the Alaska Native Service Hospital. This was done as part of the U. S. Geological Survey's program of ground-water investigations in Alaska, then under the general direction of A. N. Sayre, Chief of the Ground Water Branch of the Water Resources Division. The writer directed field work at Kotzebue.

Kotzebue is in northwestern Alaska, 60 miles north of the Arctic Circle, about 180 miles north-northeast of Nome, and 150 miles south-southeast of Point Hope. It lies near the northern end of a long, narrow peninsula forming the east shore of Kotzebue Sound; Hotham Inlet separates this peninsula from the mainland to the east.

The peninsula and the mainland are mantled by recent silt and fine sand. The area appears to be underlain by unconsolidated deposits of deltaic, estuarine, and perhaps glacial origin. The mainland to the north is bedrock. The waters surrounding the north end of the peninsula are very shallow and only slightly brackish; at times of strong flow from the mouth of the Noatak, about 5 miles across the bay, the water is fresh enough to drink.

Ordinarily, drinking water is obtained at Kotzebue from a few shallow wells occupying thawed basins in the permafrost (greatly contaminated by fuel oil and organic material), from melted ice and snow, or from somewhat distant lakes or rivers. It was hoped that a deep well there might tap a safe, dependable supply for the community of 500 or more people.

Drilling was done by the cable-tool method. The author appreciates the loan of small tools by Archie Fergusen of Kotzebue and by the U. S. Smelting and Refining Co. of Nome.

Permafrost presented some problems. Frozen ground was drilled readily enough with the periodic introduction of a little heated water or use of salt, but there is an inevitable tendency of the casing to freeze to the walls. At several times the hole at Kotzebue was thawed by steam to free the casing. Probably equally good results could have been obtained with less trouble by circulating hot water through a small-diameter line. However, the best way to prevent freezing of the casing is to drill as continuously as possible and to move the casing frequently.

The ground at Kotzebue, except for a thin stratum of gravel, is frozen to a depth of 238 feet. Beneath 19 feet of beach gravel is blue clay of marine origin, as indicated by foraminiferal content, extending to a depth of 79 feet (Fig. 1). An unfrozen stratum about 7 feet thick, seemingly another beach gravel (Fig. 2), showed promise of yielding abundant water. Statistic level was 15.9 feet below the surface (at about sea level). However, analysis showed that the water contained 26,000 ppm (parts per million) of chloride (Table 1). This salinity is appreciably greater than that of sea water (generally 18,000–20,000) ppm.

Below the unfrozen layer, frozen marine blue clay extended to 190 feet, where a gravelly layer 28 feet thick was penetrated (Fig. 1). The sample taken at 194 feet in this layer appeared to be stream laid. However, plotting of the particle-size analysis of the sample taken at 212 feet resulted in a curve typical of till (Fig. 3). Possibly the 194-foot sample also is till; it has been greatly altered by the drill, and its grain-size characteristics have been obscured. (When plotted, some samples of till from cable-tool wells in the Anchorage area look much like water-laid material.)

At 218 feet marine blue clay was again penetrated, and at 226 feet another gravel bed was

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reached. This bed, 12 feet thick, also may be glacial till.[1] A particle-size analysis of a sample taken at 234 feet shows a long diagonal slope when the cumulative percentages are plotted (Fig. 3).

All sediments between the surface and a depth of 238 feet, except the thin unfrozen stratum noted, were frozen.

At 238 feet unfrozen ground was penetrated. The heavy cable tools were thrown 10 or 12 feet upward, and the area was showered with mud and silt, owing to the release of gas, probably methane. Gas issued under pressure for about an hour; the next morning only a very small amount of gas flow could be observed.[2]

From 238 feet to 325 feet the drill penetrated silt and fine sand, possibly of continental origin. This material was saturated with somewhat salty water. Static level was 42 feet below the surface. Samples of 260 feet and 273 feet contained respectively 4680 and 5670 ppm of chloride (Table 1). The writer believed that more saline water probably would be encountered at greater depths, and the drilling was therefore discontinued. The water in the unfrozen sediments between 238 and 325 feet was unsatisfactory for the development of a producing well.

The sediments penetrated are estuarine in large part and hence were covered by bay waters above the freezing point during most of the depositional cycle. It follows, therefore, that these sediments were not frozen—or at least were not frozen throughout—until after the area was elevated above sea level.

The origin of the highly saline water between 79 and 86 feet poses an interesting question, as presumably the sediments were laid down in water less saline than the ocean of today. The writer suggests that the high salinity may have resulted from a process of fractionation by freezing of the original slightly brackish water with which the sediments were saturated. The result would have been an ice of very low chloride content and a high-chloride liquid residue. The writer proposes the following history: Freezing progressed downward from the surface. Formation of ice in the ground created a tight, impermeable mass, and excess liquid (equivalent to roughly a tenth of the volume

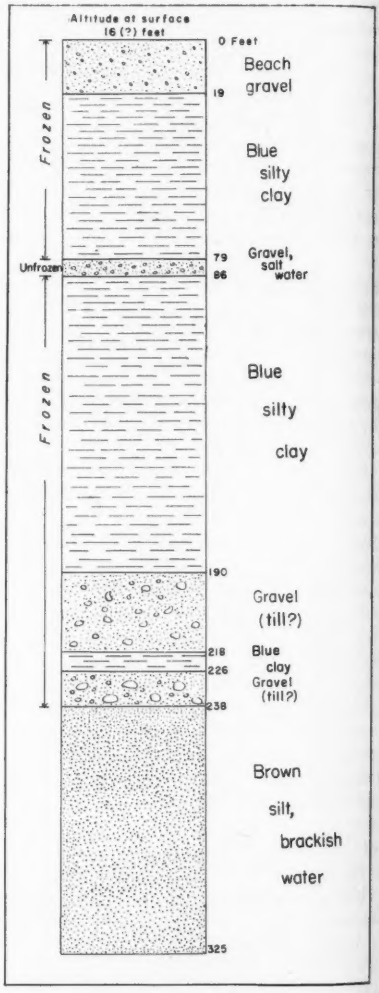

Figure 1. Log of U. S. Geological Survey test well at Kotzebue, Alaska

[1] Smith and Mertie (1930, p. 242) note that the Baird and DeLong Mountains to the north were glaciated, and from these mountains "tongues of ice extended outward into the lower country. . . ." The Kotzebue well may show evidence of the extension of an ice tongue about 50 miles from the nearest known glaciated area, probably southward from the mouth of the valley of the Noatak.

[2] Gas occurrence beneath permafrost is not uncommon. In 1948 a small accumulation was found beneath the permafrost in a Geological Survey test hole on the Farmers Loop Road, at Fairbanks. In 1952 a gas accumulation was struck beneath the permafrost on the valley flats in a water well drilled by the Army at Eielson Air Force Base, east of Fairbanks. Here the gas issued under pressure for several days.

of water) was driven downward. As the liquid with which the sediments were saturated was slightly brackish, the separation of nearly pure ice from the system increased the salinity of the unfrozen liquid below the ice front. As freezing progressed, hydrostatic pressure built up below the frozen mass, but this was relieved largely by lateral movement through the permeable gravel layer. The salt content of the originally slightly brackish water in the gravel layer was augmented by that of the high-chloride water migrating downward.

Movement of water in the gravel layer, other than that which accompanied relief of freezing pressure, must have been negligible because, if normal artesian circulation had existed, essentially fresh water would be found in the gravel bed today instead of the highly saline water, *i.e.*, flushing action would have removed high-chloride waters developed by concentration by freezing.

As the ice front approached the gravel layer, the sediments saturated with slightly brackish water probably froze at only a little below 0°C. The effective average annual air temperature was slightly lower and gradually declined appreciably below 0°C. Above the 0°C isotherm the bulk of the frozen sediments was chilled several degrees below the effective freezing point. However, as the subzero cold "migrated" downward, the salinity of the aqueous residue after fractionation by freezing finally became great enough so that the effective temperature would not produce freezing. Nevertheless, the "cold front" continued downward. Upon passing through the gravel stratum, the "cold front", or the effective temperature developing subsequent to passage downward of the "cold front", was sufficiently cold to freeze the sedi-

Table 1. Analyses of Ground-Water Samples from Test Well at Kotzebue, Alaska
Analyses by U. S. Geological Survey

	Depth (feet)			
	82	130*	260	273
Silica	9.7	5.1	42	51
Iron	68	..	..	..
Calcium	640	..	184	41
Magnesium	1890	..	..	348
Sodium and potassium	14,100	..	2850	3490
Bicarbonate	1770	221	1200	1800
Sulfate	1610	..	..	9.0
Chloride	26,200	515	4680	5670
Dissolved solids	45,300	..	..	10,600
Hardness as $CaCO_3$	9370	130	1390	1890
Specific conductance (micromhos at 25° C)	63,500	2110	15,000	17,800
pH	7.2	..	7.6	7.6

* Sample of permafrost ice

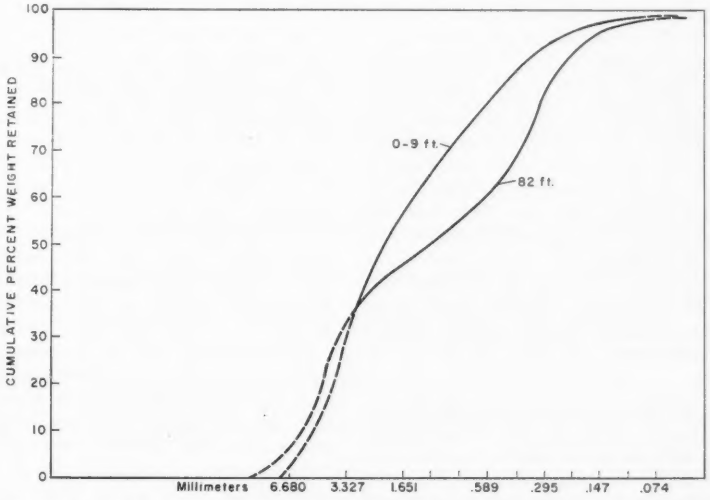

Figure 2. Grain-size analyses of upper sands penetrated in well at Kotzebue, Alaska

ments beneath (below 86 feet), because there the sediments still contained only slightly brackish water. Because of the low permeability of sediments below 86 feet, saline water migrating downward moved laterally in the gravel, and the water in the beds below the gravel bed was essentially undisturbed. Thus freezing proceeded downward to a point where the flow of heat from great depths was sufficient to prevent it.[3]

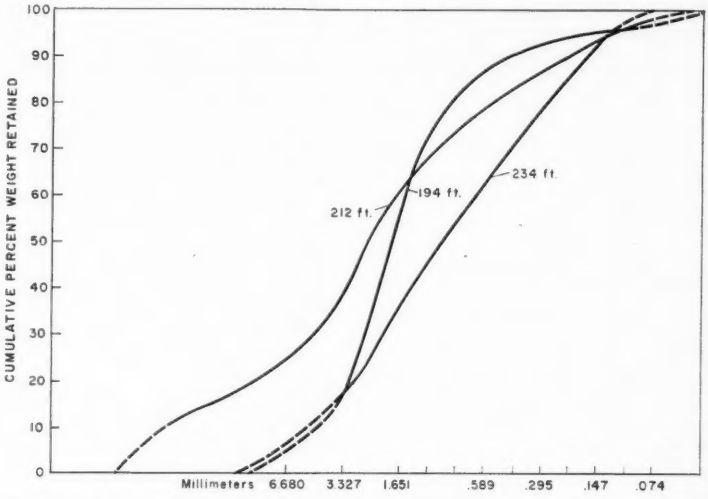

Figure 3. Grain-size analyses of deeper sands penetrated in well at Kotzebue, Alaska

One might consider it quite a coincidence that the salinity of the liquid residue after fractionation by freezing increased to the point that would prevent freezing just as the gravel bed at 79 feet was reached by the cold front. It is just as reasonable to assume that the freezing point of the residue may have been depressed to produce a nonfreezing condition after, say, 60 feet of sediments had been converted to permafrost. The cold front would then have proceeded as outlined, and the result would have been a permafrost mass with an unfrozen gravel layer overlain by about 20 feet of unfrozen silty clay. How, then, could solidification have advanced through the 20 feet of unfrozen silty clay to the gravel layer and have stopped at that point, to produce the conditions now existing?

No easy explanation presents itself, but the author suggests the following for what it may be worth. Taber's experiments on frost heaving seem to point the way, even though these experiments were performed on sediments saturated with mineral-free water and dealt with an entirely different phenomenon. Taber (1930, p. 306) shows that, in the downward freezing of cylinders of saturated clay having access to ample water below, ice layers will form and enlarge by addition of water drawn up through capillaries. On the other hand, a saturated column of sand will freeze without drawing in excess water. Taber (p. 309) emphasizes the force of molecular cohesion in the fine-grained sediments between an ice face and immediately adjacent free water particles.

Returning now to the Kotzebue problem: we have postulated a thick permanently frozen mass in which an unfrozen layer of gravel at 79 to 86 feet is overlain by about 20 feet of unfrozen clay. The salinity of the liquid in the unfrozen material is such that the freezing point of the liquid is just below that of the temperature of its environment. The writer

[3] Mr. Ted Loftus of the Fairbanks Exploration Company, Fairbanks, reported (Oral communication, August 12, 1954) that in Nome small bodies of salt water are encountered within frozen alluvium. Presumably these salt-water "lenses" likewise could have originated by the fractionation process outlined.

suggests that under these conditions, the forces of molecular cohesion in the fine-grained sediments would come into play, and that the ice faces in contact with the saline solution would continue to grow, although perhaps more slowly. However, below the contact of the clay stratum with the gravel stratum, the saline solution would be present in large voids, molecular cohesion would be negligible, and the solution would remain liquid.

According to this hypothesis, a second body of highly saline water would form below the gravel and be pushed downward below the maximum depth of freezing. Inasmuch as the samples of water taken from the hole at 260 and 273 feet were only slightly brackish, it would seem that the freezing must have extended below the depth of the lowest sample, and that the present bottom of the permafrost at 238 feet must represent a thawing back in a time of rising temperature. However, even if such thawing back has not occurred, the low salt concentration of the solution below 238 feet might well be ascribed to dilution due to normal ground-water movement in the subpermafrost zone.

It might be that the salt-water gravel represents a beach deposit formed during a glacial maximum when the sea water was more saline than average, and the fine-grained sediments above and below were deposited in fresh or only slightly brackish water during periods of rapid melting of ice. If a gravel deposit were laid down, then quickly covered and isolated hydrologically (*i.e.*, no ground-water movement occurred in the mass after burial), the observed conditions might be explained without calling upon a fractionation hypothesis. However, fractionation does occur in the Arctic, as seen in the fresh ice taken by the Eskimos from the sea as a source of drinking water, and probably as indicated in the salt-water lenses found in the alluvial plain at Nome. The entrapment of a salt-water body in a permeable lens seems less likely and might be considered a case of special pleading.

The question has been raised as to the possibility that the thawed zone reached at 238 feet at Kotzebue might be a talik (an unfrozen horizontal layer) and that frozen ground might be again encountered at greater depths. From experience in the Fairbanks area, the writer believes that the unfrozen zone at 238 feet at Kotzebue continues downward without interruption by any other frozen layer or frozen mass. In fact, the writer feels that a horizontal talik formed well down in the permafrost mass (as compared to a thawed zone between the surface winter frost and the permafrost) is an uncommon phenomenon—one that can exist only under special conditions, such as conditions of strong ground-water movement.

Melting of the upper 10 feet or so of the permafrost over a sizable area to create a reservoir of some magnitude might alleviate the present water shortage at Kotzebue (Cederstrom, 1952, p. 36). The melting would be accomplished by stripping the muskeg and admitting solar heat, circulating ground water in the thawed zone by pumping, and applying warm pond waters in summer. Undue loss of heat in winter could be prevented by building and preserving a maximum snow cover through use of snow fences. The efficiency of such a thawed reservoir would depend on depth of thaw, degree of retention of heat, and inflow of water from adjacent thawed zones lying between the winter frost and the permafrost.

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SUBAERIALLY CARVED ARCTIC SEAVALLEY UNDER A MODERN EPICONTINENTAL SEA

Abstract: A shallow seavalley, averaging 6 feet in relief, extends from the mouth of Ogotoruk Creek, northwest Alaska, for 15 miles across the floor of the Chukchi Sea to a depth of 135 feet. The seavalley is considered to be a drowned subaerial valley of Pleistocene age, which was excavated on an eustatically emerged epicontinental shelf during periods of glacially depressed sea level.

During July and August 1958, the U. S. Geological Survey conducted marine geological studies of parts of the near-shore shelf of the Chukchi Sea, Alaska, on behalf of the U. S. Atomic Energy Commission. The investigation was part of a more comprehensive geological study to select and evaluate a suitable location for a nuclear excavation test site, which was finally chosen at the mouth of Ogotoruk Creek, northwest Alaska (Fig. 1).

This paper calls attention to a shallow but long seavalley discovered during the investigation. A more comprehensive report of the bathymetry, oceanography, and marine geology will appear in the forthcoming Proceedings of the First International Symposium on Arctic Geology (Scholl and Sainsbury, in press).

The authors conducted marine geologic studies from an outboard-powered skiff along a section of coast line 20 miles long and 5 miles wide by use of lead-line soundings and bottom grabs. Holmes and Narver, Inc., of Los Angeles, California, subsequently collected and compiled sonic data in the central part of the area. Bathymetric information seaward of the 60-foot contour was taken with acoustic equipment by the icebreaker, USS Burton Island. Sextant fixes, ranging angles (simultaneous sighting to sea by two transits), and radar fixes to coastal reference points were used, respectively, by the three groups. All the sounding data were available to the writers, and approximately 3000 soundings were used in constructing the 340 square miles of submarine topography shown on Figure 1. Sounding density in the area ranges from 1.4 to 270 per square mile and averages 8.4.

The Chukchi Sea, which has a maximum depth of about 180 feet, has been described as one of the flattest areas on earth (Buffington, Carsola, and Dietz, 1950). The seavalley described, here named Ogotoruk Seavalley, heads in 30 feet of water approximately a quarter of a mile off the mouth of Ogotoruk Creek. The axis of the valley is sinuous but trends generally southeastward; a number of tributary valleys enter near the head of the seavalley. Ogotoruk Seavalley has a maximum relief of 15 feet and an average relief near 6 feet.

Near the shore the seavalley is approximately 2 miles wide and contains a number of knolls rising as much as 5 feet above the valley floor. The gradient of the seavalley in this part is 7 minutes. At depths greater than 85 feet, the seavalley narrows to a single channel about half a mile wide, and then gradually widens to merge with the shelf of the Chukchi Sea at a depth of 135 feet. The gradient of the seavalley in the outer 5 miles is about 2 minutes.

Ogotoruk Seavalley is similar to other submarine valleys in (1) its sinuous course, (2) its smooth gradient, (3) its tributaries entering at grade, and (4) in having been cut in bedrock. The authors found bedrock along the sides and bottom of the inner part of the valley, and divers from the USS Burton Island found it within a few hundred yards of the shore. Ogoturuk Seavalley differs from other seavalleys in having (1) a broad, although somewhat hummocky, flat profile; (2) a shallow relief; and (3) a gradient nearly 300 times lower than other typical submarine valleys (Kuenen, 1950, p. 487).

Bottom samples, visual observations in clear water, and reports of divers indicate that steeply dipping bedrock crops out over much of the near-shore shelf shoreward of the 50-foot contour (inner 4–5 miles). The flat bedrock surface of the near-shore shelf abuts precipitous sea cliffs as much as 800 feet high, and the near-

Geological Society of America Bulletin, v. 72, p. 1433–1436, 1 fig., September 1961

shore shelf in this area is interpreted as a wave-planed platform bevelled across rocks of extremely variable hardness.

Ogotoruk Seavalley is cut into this bedrock platform and is interpreted therefore as a drowned subaerial valley eroded by Ogotoruk Creek on an emergent wave-planed shelf. The evidence for this interpretation is (1) the alignment of the seavalley with the valley of Ogotoruk Creek, (2) the similar width and morphology of the two valleys, and (3) the fact that both valleys have similar gradients, with a progressive decrease in gradient seaward.

The planation of the near-shore shelf may have been accomplished at any time since the formation of the Chukchi Sea, but a wave-planed terrace somewhat higher than the modern platform is exposed intermittently along the coast and can be correlated with the higher Sangamon sea, of late Pleistocene age, showing that wave-planed terraces or platforms were cut prior to the present stand of sea level. The near-shore shelf may, therefore, represent more than one period of marine planation. The time of cutting of the seavalley cannot be precisely determined owing to the lack of a clear record of pre-Sangamon events. Without question, however, Ogotoruk Seavalley was occu-

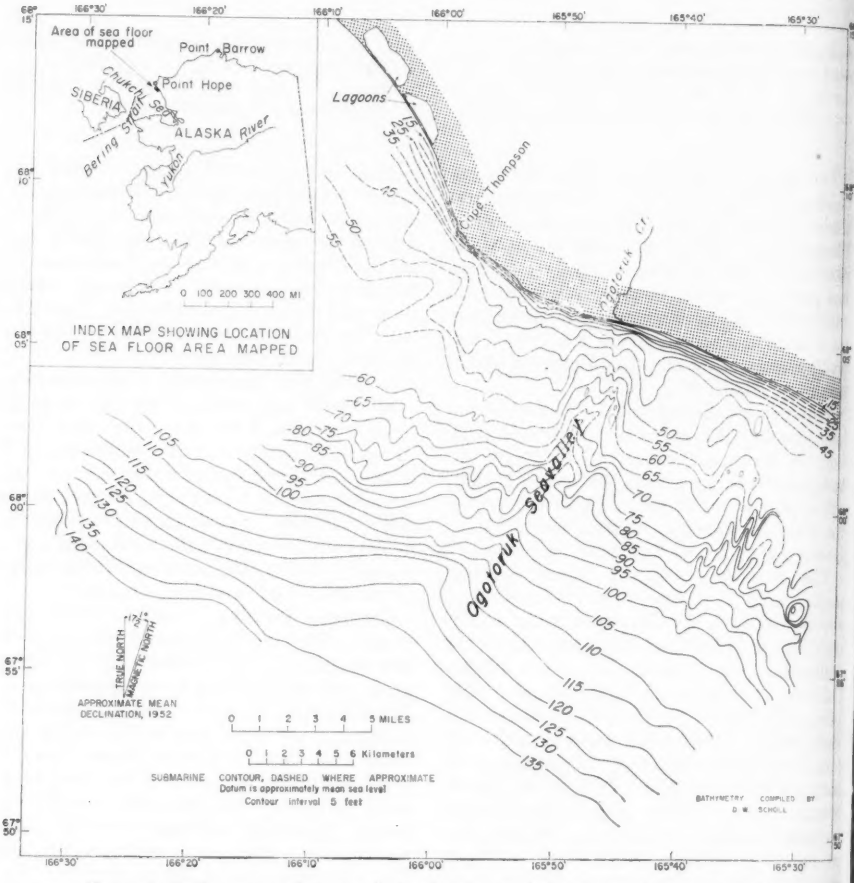

Figure 1. Bathymetry off Ogotoruk Creek area, Chukchi Sea, northwestern Alaska

pied by Ogotoruk Creek during the period of low sea level of Wisconsin time, and erosion of the seavalley may have been accomplished principally during the Wisconsin.

The conclusion that Ogotoruk Seavalley was carved during glacial ages requires, during times of maximum glaciation, a frost-free season in the Arctic, similar to that at present. Such a frost-free season during glacial ages is compatible with, but not proof of, Ewing and Donn's (1956) hypothesis that the Arctic Ocean was free of ice at times of Pleistocene glaciation.

Because Ogotoruk Seavalley was formed by subaerial erosion, it is apparent that other streams, especially the larger ones, probably also carved valleys on the emergent Chukchi platform, and future studies may well delineate a complex Pleistocene drainage etched onto the floor of the Chukchi Sea.

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AGE OF THE GLEN CANYON GROUP (TRIASSIC AND JURASSIC) ON THE COLORADO PLATEAU

Abstract: The paleontologic and stratigraphic evidence for age designations of the Glen Canyon group has been reappraised. This evidence suggests that the Kayenta formation is of Late Triassic(?) rather than Early Jurassic(?) age, and that the lower part of the Navajo sandstone is of Late Triassic(?) rather than Jurassic(?) age.

Age designations for the formations of the Glen Canyon group of the Colorado Plateau as recognized by the U. S. Geological Survey were, in descending order:

Navajo sandstone—Jurassic and Jurassic(?)
Kayenta formation—Early Jurassic(?)
Moenave formation—Late Triassic(?)
Wingate sandstone—Late Triassic

A recent appraisal of paleontologic and stratigraphic evidence by a group of geologists[1] of the U. S. Geological Survey at Denver resulted in the proposal that the Kayenta formation and the lower part of the Navajo sandstone be assigned a Late Triassic(?) age.

The Moenave formation, which as currently recognized consists of the Dinosaur Canyon sandstone member at the base and the Springdale sandstone member at the top, is now questionably referred to Late Triassic time. It has yielded the holostean fish *Semionotus* (*Lepidotus*) *walcotti*, *S. kanabensis*, and *Semionotus* cf. *S. gigas* (Eastman, 1917; Hesse, 1935; Schaeffer and Dunkle, 1950). The crocodilian *Protosuchus richardsoni* (Brown, 1933; 1934; Colbert and Mook, 1951), the only member of the suborder Protosuchia known outside of South Africa, is present in the Dinosaur Canyon member 24 miles south of Moenave, Arizona (Fig. 1). Most authorities accept all these species as being of Triassic age.

Protosuchus richardsoni and a new genus of tritylodont reptile are present from 8 to 10 feet below the top of the Kayenta formation on Comb Ridge 6 miles east of the type locality near Kayenta, Arizona (Fig. 1). Not only does the presence here of *Protosuchus* suggest an age for the Kayenta essentially the same as that of the Moenave, but its association with the tritylodont also suggests a correlation of the Kayenta and Moenave with the Stormberg series of South Africa. The Stormberg series contains *Tritylodon longaevus*, related to the tritylodont found in the Kayenta, associated with the protosuchian species *Notochampsa istedena* and *Erythrochampsa longipes*, and with species of *Semionotus*. The Kayenta tritylodont is more closely related to *Tritylodon* and to *Bienotherium*, a tritylodont described by Young (1947) from the Lufeng series (Keuper) of China, than to the known Rhaetic to Early Jurassic *Oligokyphus*, or to the Middle Jurassic *Stereognathus*, the last survivor of the tritylodonts. Watson (1942) has made a careful evaluation of the upper Stormberg fauna and its probable age; he correlated it with the Keuper (Late Triassic, older than the latest Triassic Rhaetic). Haughton (1924) has made an exhaustive study of Stormberg stratigraphy and also has studied the Stormberg fauna; he too correlated the Stormberg with the Keuper. There have been some, however, who believed the Stormberg is of Rhaetic or of Early Jurassic age (Broom, 1904). We believe it is fair to say that today most authorities favor a Keuper age.

A carnivorous dinosaur, "*Megalosaurus*" *wetherilli*, has been described by Welles (1954) from beds assigned to the lower part of the Kayenta. Welles believed the dinosaur to be Jurassic because of the relative proportions of the limb bones as compared to those of other Megalosaurs. However, Swinton said (1955, p. 134): ". . . the main elements of the skull, the scapula, and the pelvic bones are not those of typical Megalosaurs . . . it is not *Megalosaurus*."

[1] F. W. Cater, Jr., L. C. Craig, J. H. Irwin, G. E. Lewis, S. W. Lohman, E. D. McKee, F. G. Poole, J. D. Strobell, Jr., R. F. Wilson, and J. C. Wright

Geological Society of America Bulletin, v. 72, p. 1437–1440, 1 fig., September 1961

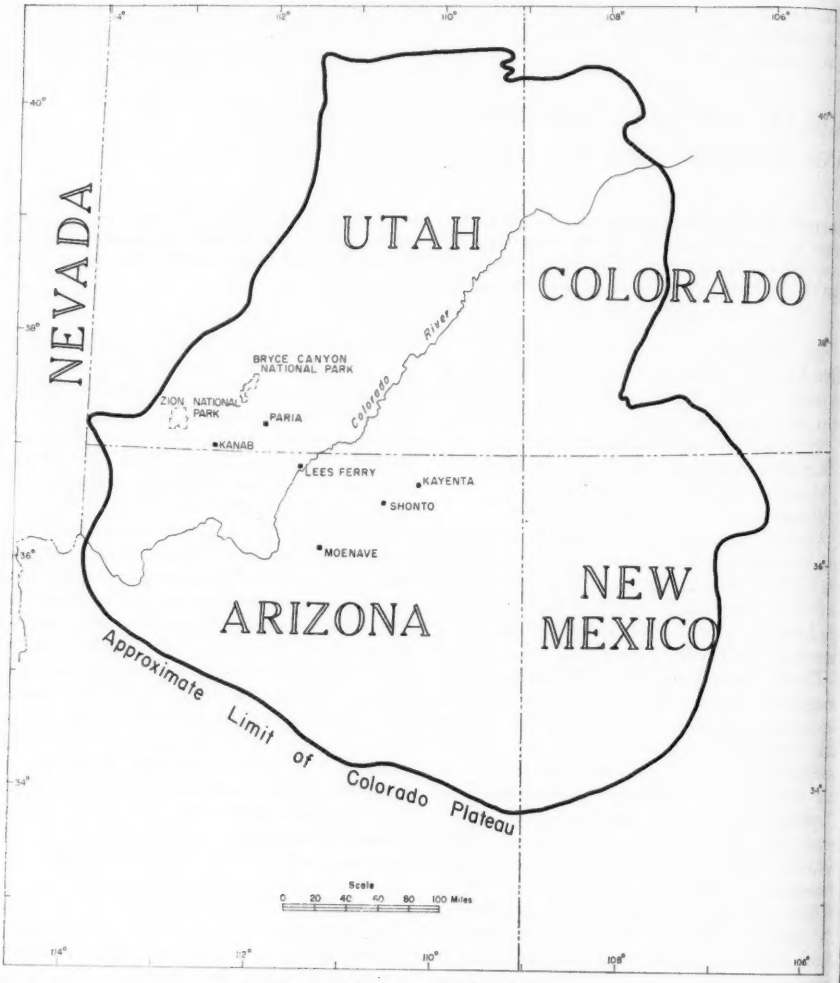

Figure 1. Index map of Colorado Plateau

It would therefore seem that the comparisons of the limb proportions of this dinosaur to the limb proportions of typical Megalosaurs are not applicable for age determinations. Whatever the genus may be, it should be closely compared to the carnivorous dinosaurs found in the Lufeng with *Bienotherium*.

Invertebrate remains also have been reported from the Kayenta formation. They consist of a species of fresh-water pelecypod, *Unio* sp., which is long ranging and not diagnostic of age (Baker, 1933), and two species of fresh-water gastropods, *Valvata gregorii* and *Lymnaea hopii*, which Yen (1951) assigned to the Late Triassic. Therefore the evidence, although not conclusive, strongly suggests that the Kayenta formation is of Late Triassic age.

Finally, the age and relations of the overlying Navajo sandstone should be re-evaluated in the light of recent field work, and the fauna

should be reconsidered. Harshbarger, Repenning, and Irwin (1957, p. 28) reported collections identified by J. B. Reeside, Jr., R. W. Brown, R. E. Peck, and I. G. Sohn. These collections, from near Shonto, Arizona (Fig. 1), include nondiagnostic bivalved Crustacea referred to *Lioestheria* aff. *L. ovata* (Lea)—formerly referred to *Cyzicus*—and to a genus similar to *Candona*, fragmentary fern leaves and wood, and the very long-ranging plant *Equisetum*. These invertebrates and plant remains are of no help in age determination, but Reeside commented that "the association is much like that in the Newark Triassic." The only other known fossils from the Navajo are two genera of carnivorous dinosaurs also from near Shonto: *Segisaurus*, from the middle part of the Navajo "500 feet above its base," was described by Camp (1936, p. 39, 52) who stated that "it could be placed in either the Triassic or the Jurassic." The other was referred by Brady (1935, p. 210, 215; 1936, p. 150) to *Ammosaurus* sp., a genus characteristic of the Newark group of Late Triassic age of the Connecticut Valley; it was collected from "80 to 100 feet below the top of the formation" and only about 0.1 mile from the nearest outcrop of the overlying Carmel. In short, the scanty evidence suggests a Late Triassic age for much of the Navajo.

The stratigraphic relations support the paleontologic evidence. We believe that the Springdale sandstone member of the Moenave formation correlates, at least in part, with the type section of the Kayenta formation near Kayenta, Arizona. This correlation seems valid because the Springdale and type Kayenta are lithologically identical, occupy the same stratigraphic interval, and appear to have the same source and transport direction. Harshbarger, Repenning, and Irwin (1957, Pl. 2) showed that the Springdale sandstone member grades into and is indistinguishable from the Kayenta formation about 10 miles west of Kayenta. There is a similar gradation in the Lees Ferry and Paria areas (Fig. 1). These stratigraphic relations indicate that the Moenave and Kayenta formations are partial stratigraphic equivalents.

Large-scale intertonguing between the Kayenta formation and the overlying Navajo sandstone is present in northern Arizona and southwestern Utah (Harshbarger, Repenning, and Irwin, 1957, p. 18). Averitt, in the Cedar City area 20 miles north of Zion National Park, and Wilson, in the Kanab area (Fig. 1), also have demonstrated that the lower part of the Navajo sandstone intertongues with the Kayenta formation (Averitt, *et al.*, 1955, p. 2520–2523). The intertonguing between the Kayenta and Navajo occurs through an interval of as much as 700 feet and extends laterally for at least 100 miles. Accordingly the lower part of the Navajo, which intertongues with the Kayenta, should be considered to have the same age as the Kayenta.

Intertonguing between the upper part of the Navajo sandstone and the fossiliferous Carmel formation of Middle and Late Jurassic age in southwestern Utah has been demonstrated through recent field work by D. A. Phoenix and J. C. Wright of the U. S. Geological Survey. A prominent tongue of the Navajo about 200 feet thick near Lees Ferry, where it separated from the main body of the formation, has been mapped by D. A. Phoenix (1961, written communication) and traced northwestward about 70 miles along the outcrop to the base of the Pink Cliffs near Bryce Canyon National Park (Fig. 1).

A still lower part of the Navajo sandstone, named the Temple Cap member by Gregory (1950, p. 89), also extends as a tongue into the Carmel formation. It is about 150 feet thick where it separates from the rest of the formation north of Kanab and it can be traced nearly 30 miles to the western part of Zion National Park (J. C. Wright and F. G. Poole, written communication, July 27, 1960). The part of the Navajo sandstone that intertongues with the Carmel formation, therefore, can be considered of definite Jurassic age.

The stratigraphic evidence indicates the following: the Moenave and Kayenta are partial rock equivalents, the Kayenta and the lower part of the Navajo are partial rock equivalents, and the upper part of the Navajo and the lower part of the Carmel are partial rock equivalents. Equivalent, intertonguing parts of any two of these formations must have a similar age, at least in the area of intertonguing.

In light of the paleontologic and stratigraphic evidence presented here, the Geologic Names Committee of the U. S. Geological Survey has approved the following age assignments for the formations of the Glen Canyon group, in descending order:

Navajo sandstone—Late Triassic(?) and Jurassic

Kayenta formation—Late Triassic(?)
Moenave formation—Late Triassic(?)
Wingate sandstone—Late Triassic

We would have suggested reassignment of the Kayenta and Moenave formations and also the lower part of the Navajo sandstone to the Triassic without query, except that: (1) the formations are unfossiliferous in most areas; (2) there are still some differences of opinion regarding the age of some of the fossils.

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FRANCIS BIRCH

ROLE OF FLUID PRESSURE IN MECHANICS OF OVERTHRUST FAULTING: DISCUSSION

The paper by Hubbert and Rubey (1959) on the mechanics of overthrusting brings out very clearly some of the problems that arise in the attempt to account for the displacement of a relatively intact thrust sheet having a breadth of the order of 100 km. The statement (p. 126–127) that "for the conditions assumed, the pushing of a thrust block, whose length is of the order of 30 km or more, along a horizontal surface appears to be a mechanical impossibility" is undoubtedly correct if by the "conditions assumed" we understand that the strength of the block (τ_0) is about 200 bars and the coefficient of sliding friction on the base is 0.577 or greater. The way out of this dilemma proposed by Hubbert and Rubey is to reduce the coefficient of friction by the introduction of abnormally high pore pressure; the calculations then lead to much greater lengths of thrust blocks (Table 1, p. 144), which might, indeed, be increased by another factor of 10 for a pore pressure factor of 0.99 (versus 0.9), 100 for 0.999, and so on. This feature of their solution is examined, and several inconsistencies are pointed out. It is, of course, clear that the model adopted for this analysis is extremely simplified, and that many factors that might be of some consequence (temperature, time, recrystallization, different criteria of rupture, and so on) are necessarily ignored. One might thus reasonably question the necessity for perfect consistency and rigor in dealing with a model that departs so far from geological plausibility; on the other hand, and this is the point of view from which this discussion is written, it may be argued that the best starting point for dealing with the complex reality is a self-consistent solution even for a highly simplified model.

In the Hubbert-Rubey analysis, the thrust block is treated as a homogeneous rectangular block resting on a plane surface, pushed from one end or allowed to slide on an incline. The question to be decided is whether a block of given dimensions and properties will first slide on the plane of the base or rupture and slide on some internal surface. Two independent conditions must be introduced: (1) a criterion for sliding on the base, and (2) a criterion of rupture for the interior of the block. The Coulomb-Navier condition is adopted as the criterion of rupture; this is represented in the (two-dimensional) Mohr diagram by a line (or pair of lines) having the equation, $\tau = \tau_0 + \sigma \tan \phi$, where ϕ is the "angle of internal friction", or in a diagram of normal stress by a line, $\sigma_1 = a + b\sigma_3$, where σ_1 and σ_3 are principal stresses. Experiments with rocks suggest a value for ϕ of about 30°, or $\tan \phi = 0.577$.

This number, 0.577, is also taken as the coefficient of friction on the base for the "dry" case. This reveals a basic ambiguity regarding the relationship of the thrust block to the underlying medium. If the block is merely a projection on an underlying layer having the same properties, then neglect of τ_0 in the condition for sliding (equations 11 and 94) is no more justified for the base than for any other internal surface; furthermore, fracture would never follow the plane of the base, but would curve upward as suggested by Hubbert's (1951) experiments and Hafner's (1951) calculations. On the other hand, if the thrust block is considered to be separate from the underlying layer, then we need consider only the coefficient of sliding friction between the two bodies, and this has nothing to do with the "angle of internal friction" related to the rupture condition.

When the normal stress is high, as beneath a thick "dry" layer, the neglect of τ_0, of the order of a few hundred bars, makes no great difference. With the introduction of pore pressure, however, the criterion of rupture is transformed to $\tau = \tau_0 + (1 - \lambda)S \tan \phi$ where S is the total normal stress, and λ is the pore-pressure coefficient. Since λ approaches 1 as pore pressure approaches rock pressure, the second term approaches zero. Neglect of τ_0 by comparison with the vanishing second term for the sliding criterion on the base, while retaining τ_0 in the criterion for internal rupture, is

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the key to the great lengths of overthrust shown in Table 1 (p. 144). In effect, the neglect of τ_0 in the condition for slip on the base leads to a coefficient of friction on the base which may become zero. But in this case pore pressure reduces friction on all planes and leads to a system devoid of shear strength for λ approaching 1; if τ_0 is consistently retained, this becomes also the minimum shear stress for the slip on the base, and the coefficient of friction on the base remains finite.

Several other inconsistencies appear in the calculations of length of thrust block. The stresses in a block loaded as shown in Figures 5, 6, 19, and 20 must include shear stresses on the vertical sides; the systems of forces shown in these figures are not in equilibrium. Because of the existence of the components τ_{xz}, the normal components are not principal stresses, and their substitution in the criterion of rupture, taken in the form $\sigma_1 = a + b\sigma_3$, leads to an overestimate of the length of block that can be pushed with a given coefficient of friction on the base and given τ_0. This error diminishes as the ratio of length to height of block increases, but, for the assumed conditions, internal rupture will *always* take place before sliding on the base for the dimensions and numbers given by Hubbert and Rubey. This reinforces their conclusion with respect to the "dry" case, but it also undermines the apparent value of pore pressure as a mechanism for producing long thrust blocks.

The face free from lateral pressure is also a place of weakness; with τ_0 taken as 200 bars (p. 126), failure at the base of this cliff begins when the height reaches about 3 km. For consistency, if greater thicknesses are to be considered, either τ_0 should be correspondingly increased, and this would increase the length of thrust block that could be moved in one piece, or there should be a distribution of normal pressure on the free face, which would prevent failure at this place.

The Hubbert-Rubey calculation for a block 5 km thick gives 18.4 km for the maximum length, and a more rigorous calculation gives 15.6 km, but this block has already failed at the free face. If we require that the strength be adequate to hold up a vertical free face, then the maximum length for a 5-km block turns out to be 28.6 km ($a = 1240$ bars, $\tau_0 = 360$ bars).

No one supposes, of course, that a vertical cliff of such height is necessary for overthrusting; it occurs in the model only because the most favorable conditions for sliding seem to require a free face. We cannot give much meaning, however, to a solution corresponding to failure at the loaded face if failure has already occurred at the free face.

Conditions at a free face become especially anomalous when there is high pore pressure. The condition that the total normal stress S_{xx} be zero implies, since $S_{xx} = \sigma_{xx} + p$, that the stress σ_{xx} in the solid portion becomes tensile and equal in magnitude to the pore pressure p. This not only departs from realism, but insures failure at the free face for heights greater than 1.1 km ($a = 700$ bars, $\lambda = 0.9$) even without a lateral push. If τ_0 is neglected in the interior of the block, as on the base, then no free vertical face is possible; this is the case of the sand pile, for which a free face approaches a slope having the "angle of repose".

In order to account for thrust blocks of greater length, it is necessary to introduce smaller coefficients of friction on the base. The Hubbert-Rubey theory proposes that with sufficiently high pore pressure the coefficient of friction can be made to vanish on the base, while remaining finite elsewhere. The inconsistency of this treatment has already been mentioned. The mechanism of pore pressure can be helpful, however, if the high pore pressure is confined in the *underlying* layer, rather than in the thrust block. In this case, as pore pressure increases, the condition of failure of the underlying medium approaches the Tresca condition of maximum shear stress, equal to the τ_0 of the underlying medium. This may be no more than 100 bars or so, even for relatively strong rocks such as the Berea sandstone (p. 142). For a 5-km layer, the normal stress on the base is 1130 bars (with $\rho = 2.31$) so that slip for a shear stress of 113 bars corresponds to a coefficient of friction of 0.1 and a possible length of thrust block of the order of 100 km. This would also be the case, without pore pressure, if the underlying medium has a maximum shear strength of the order of 100 bars. The essential condition for overthrusting is the combination of a strong thrust block with a weak underlying layer; abnormal pore pressure may be effective in weakening the underlying layer, but it will not help if it weakens the thrust block as well.

The figure 0.577, or tan 30°, has no special significance as a coefficient of sliding friction; it comes into the discussion by way of the criterion of rupture. Much lower coefficients are possible for heavy blocks moving on weak

layers. In many soft materials, the shear stress at yield is little affected by normal stress; when these are used as lubricants, the coefficient of friction becomes approximately reciprocal to the normal load. In a familiar laboratory device, hardened steel surfaces are separated by thin lead foil, and moved under normal pressures of the order of 10 kilobars with effective coefficients of friction as low as 0.005.

A word on the demonstrations with the beer can and the concrete block may be in order. The reduction of sliding friction by the presence of a pressurized liquid film which lightens or prevents contact between solid surfaces is unquestionable. This, however, is not analogous to pore pressure; tests of rocks and concrete show that a finite shear strength remains even when the pore pressure equals the external pressure (in contradiction to the assumption of equation 94, p. 143). There is no pore pressure in the beer can, nor in the concrete model. If, however, there is a continuous liquid layer which completely separates the solid surfaces, then frictional resistance must approach some low value determined by the viscosity and dimensions of the liquid layer and the rate of displacement of the solids. The coefficients of internal friction of the solids play no part in this effect.

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ROLE OF FLUID PRESSURE IN MECHANICS OF OVERTHRUST FAULTING, I. MECHANICS OF FLUID-FILLED POROUS SOLIDS AND ITS APPLICATION TO OVERTHRUST FAULTING: REPLY TO DISCUSSION BY FRANCIS BIRCH

Introduction

The authors have studied with great interest Prof. Francis Birch's criticism of their hypothesis (Hubbert and Rubey, 1959) that the paradox of overthrust faulting can largely be resolved if account is taken of the buoyant forces and other mechanical effects that are produced in the rocks by the pressure of the water with which their pore spaces are filled. They find themselves in substantial agreement with a number of the points he has made, but they still are unable to accept his principal conclusions for the following reasons:

(1) They think that he is wrong in his insistence upon retaining the term τ_0 of their equation

$$\tau = \tau_0 + (1 - \lambda)\, S \tan \phi$$

in computing the maximum length of an overthrust block that can be pushed over a horizontal surface.

(2) They think that he misses an essential element of the mechanics of the phenomena considered when he refers repeatedly to the reduction of the coefficient of friction which he presumes to be brought about by an increase of the pressure of the pore water.

(3) They think that his criterion of self-consistency in what he refers to as the authors' "extremely simplified" model, "which departs so far from geological plausibility," is not germane and that his criticisms based on this are largely misdirected.

(4) They think that his interpretation of the concrete-block and beer-can experiments is mechanically erroneous and that his denial of their geological significance is unwarranted.

These questions will be considered in the order enumerated.

1. Neglect of τ_0.

The authors' equation,

$$\tau = \tau_0 + \sigma \tan \phi\,, \qquad (1)$$

(Hubbert and Rubey, 1959, Equation 3, p. 123, Fig. 2, p. 124) is the equation of the Mohr envelope for rocks in the brittle domain that have been tested to failure under triaxial stress. In this equation τ is the shear stress along the surface of failure at the time the specimen fails, σ is the normal stress acting across this surface, and τ_0 and ϕ are mechanical constants for the rock being tested. From the equation, the internal-cohesion term τ_0 is the shear stress at which the rock will fail if the normal stress is zero, and ϕ, known as the angle of internal friction, is the angle of slope of the Mohr envelope. The angle ϕ has an average value for a wide variety of rocks of about 30°.

The above equation applies to the "dry-rock" case where the tests are carried out on jacketed specimens with zero internal pore pressure. The authors showed (p. 138–142) that when the same rocks are tested while subjected simultaneously to an internal pore pressure p, the above equation is still valid provided that the normal stress σ is interpreted to be the *effective normal stress*,

$$\sigma = S - p\,, \qquad (2)$$

as defined by Terzaghi, where S is the normal component of total stress across the surface of slip.

Then, letting p/S be represented by λ, we obtain

$$\sigma = (1 - \lambda)\, S\,, \qquad (3)$$

which, when inserted into equation (1), gives for the critical shear stress at the time of failure

$$\tau = \tau_0 + (1 - \lambda)\, S \tan \phi\,. \qquad (4)$$

The question of present interest is whether or not the term τ_0, which from test data on sedimentary rocks was found to have a mean value of about 200 bars, can or cannot be neglected when the surface of slip is that of a large overthrust fault. Concerning this ques-

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tion, the authors stated in the original paper (p. 125):

"Referring to equation (3) [$\tau = \tau_0 + \sigma \tan \phi$], it will be seen that the total shear stress required to initiate a fracture on a small specimen is the sum of the initial shear strength τ_0 and the shear resistance to sliding. Once the fracture is produced, only the latter resistance remains. On an area as large as the surface of a regional fault, it appears to be most likely that the fracture will be propagated as a dislocation. If so, the area over which the stress τ_0 would be involved at any one time should comprise but a minute fraction of the total area of the fault surface. We shall accordingly assume that the effect of τ_0 is negligible."

This is opposed by Birch in the following statement:

". . . Neglect of τ_0 by comparison with the vanishing second term [of equation (4) herein] for the sliding criterion on the base, while retaining τ_0 in the criterion for internal rupture, is the key to the great lengths of overthrust shown in Table 1 (p. 144). In effect, the neglect of τ_0 in the condition for slip on the base leads to a coefficient of friction on the base which may become zero."

We must accordingly consider two questions: (1) Is the neglect of τ_0 under any condition permissible? (2) If τ_0 may be neglected on the base of the fault block, is it permissible to retain it in the interior of the block?

It is the view of the authors that both these questions may be answered in the affirmative. In the first place, τ_0 is only significant *before a fracture is started*, but in that domain it must be taken into account in estimating the maximum stress that the rock can sustain. After the fracture has been formed τ_0 vanishes to zero along the fracture, *but not elsewhere*. In this state a maximum stress based upon a nonzero τ_0 can be applied to the block which is then opposed by a shear stress of ordinary sliding friction along its base.

On this point Professor Birch asserts:

". . . if the thrust block is considered to be separate from the underlying layer, then we need consider only the coefficient of sliding friction between the two bodies, and this has nothing to do with the 'angle of internal friction' related to the rupture condition."

If the authors agreed with the above statement they could simply write

$$\tau = (1 - \lambda)\, S \tan \mu \qquad (5)$$

for the stress required to produce sliding, where μ would be the angle of sliding friction, and $\tan \mu$ the corresponding coefficient. Then, since the experimental values of μ for rocks are about 30°—the same as ϕ—equation (5) would give the same shear stress as equation (4) with τ_0 eliminated. The authors do not agree with the statement, however, and see no reason for replacing ϕ by another symbol for an angle *having the same numerical value and the same physical significance*. That the principal difference between "internal" and sliding friction for brittle materials is the presence or absence of an initial cohesion indicated by τ_0 is shown by tests on loose sand. Here τ_0 is zero, but the angle of sliding friction as determined by the angle of repose, and the angle of internal friction, obtained by tests involving slip on internal surfaces, are the same and are merely two equivalent manifestations of the same basic phenomenon.

Let us consider now how a fault rupture comes into existence. We will accept as an empirical fact that overthrust-fault surfaces are nearly horizontal. If the fault is initiated and the thrust block is moved by a compressive stress applied along its rearward edge, then the stress must have its greatest intensity in that region. As the stress is built up it reaches the critical value for rupture and a fracture is started—not multiple fractures throughout the rock, but one particular fracture, that of the fault itself. Then as the fracture advances the part of the block to the rear of the leading edge of the fracture suffers only the frictional resistance to sliding given by equation (4), above, with the term τ_0 deleted, which may, as the authors showed and the critic agrees, be made to approach zero.

Under these circumstances the very large force that may be exerted on the rear edge of the block is principally opposed by a stress concentration localized in the vicinity of the leading edge of the fracture. This can easily be greater than τ_0, and the fracture can be propagated as long as such a thrust from the rear can be transmitted to its leading edge.

The difference between the total force required to move a block of a given size when the fracture is propagated in this manner and that which would be required if τ_0 were effective simultaneously over the total basal area of the block can accordingly be quite large. Since the force is the product of the stress by the area, if A is the area of a longitudinal strip of the fault surface and F_A the force required to overcome τ_0, assumed to be effective over all of A simultaneously, then

$$F_A = \tau_0 A\,. \qquad (6)$$

On the other hand, if τ_0 is effective only over a small area ΔA, then

$$F_{\Delta A} = \tau_0 \Delta A \tag{7}$$

and

$$F_{\Delta A}/F_A = \Delta A/A\,, \tag{8}$$

which, as A becomes large, approaches the limit zero.

A good analogy to these two cases is afforded by tearing a piece of cloth. Consider a piece of cloth 1 m wide with a tensile strength of, say, 5 kgm weight per cm of width. If uniform tensile stress were applied to the whole width of the cloth the force required to tear it would be 500 kgm weight, or half a ton. On the other hand, if the stress is applied initially at one edge only, a tear can be started at that point and then propagated across the cloth by a force that can easily be applied with one's two hands, say, 2–5 kgm weight.

Since fault fractures do not originate instantaneously over the whole of their ultimate extent, but have some local place of origin and then propagate edgewise, it is quite as erroneous to compute the force required to move the block by assuming that τ_0 is effective at any one time over the whole of the fault surface as it would be to compute the force required to tear a piece of cloth in the conventional manner by assuming that the tensile strength is simultaneously effective over the whole width.

2. Reduction of Coefficient of Friction

In one of the passages quoted above Birch states:

"... In effect, the neglect of τ_0 in the condition for slip on the base leads to a coefficient of friction on the base which may become zero."

Later, he remarks:

"In order to account for thrust blocks of greater length, it is necessary to introduce smaller coefficients of friction on the base. The Hubbert-Rubey theory proposes that with sufficiently high pore pressure the coefficient of friction can be made to vanish on the base while remaining finite elsewhere."

This is a thesis to which he consistently adheres throughout his discussion, notwithstanding the fact that in their abstract the authors explicitly stated (p. 115):

"From equations (4) and (6) it follows that, *without changing the coefficient of friction tan* ϕ, the critical value of the shearing stress can be made arbitrarily small simply by increasing the fluid pressure p." [Italics added.]

Equations (4) and (6) referred to are equivalent to equations (1) to (4) herein with the term τ_0 deleted, which leaves

$$\begin{aligned} \tau = \sigma \tan \phi &= (S - p) \tan \phi \\ &= (1 - \lambda)\, S \tan \phi \end{aligned} \tag{9}$$

as the equation for the shear stress required to slide one block over the other.

The validity of this equation was established experimentally by the concrete-block experiment described in the original paper (p. 158–161). Since ϕ is an experimentally determined constant for the given materials, and tan ϕ is the coefficient of friction *by definition*, on what grounds can the critic claim that an increase in λ produces a corresponding decrease in the coefficient of friction? Actually, as the equation unambiguously states, and as the authors pointed out originally, without changing either the coefficient of friction tan ϕ, or the total stress S, but merely by reducing the normal component of stress σ by increasing λ, the shear stress required to produce sliding can be made arbitrarily small.

This failure to distinguish between two entirely different ways of reducing the shearing stress required to produce sliding—by the reduction of the normal stress σ, or by the reduction of the coefficient of friction tan ϕ—has led the critic unnecessarily to seek ways of reducing the latter. The most obvious, of course, is by lubrication, whereby the shearing stress required to slide one rigid plate with respect to another can be reduced if the two plates are separated by a layer of weak plastic material. As an example, the critic has cited the use of lead foil between hardened steel surfaces, where, with a normal stress of 10 kilobars, effective coefficients of friction as low as 0.005 can be achieved.

The principle here invoked is valid without question. If a suitable lubricant among the rocks involved in overthrust faulting can be found, and if it can be shown that the fault always follows such a lubricating layer, then this would afford strong support of this particular method of reducing the force required to move the block. The lubrication mechanism has in fact been proposed intermittently for the last three-quarters of a century, but upon measurement the rocks involved in overthrust faulting rarely have the required rheological properties. In fact, on the basis of hundreds of

laboratory tests under triaxial stress they exhibit predominantly the behavior described by equations (1) and (4) herein, with a friction angle of about 30°. This largely precludes the lubrication hypothesis.

What the authors have shown, on the other hand, is that there exists another mechanism for reducing the shear stress required for sliding, which is entirely compatible with the observed coefficients of friction and the observed strength of the rocks involved in overthrust faulting. Hence a low coefficient of friction, or a weak plastic layer, while not excluded as a possibility, is no longer a logical necessity.

3. Inconsistent Analysis of Oversimplified Model

Aside from the fundamental mechanical questions just discussed, Professor Birch's principal criticism has been directed at what he refers to as the authors' "model." He concedes that a simplified model of the more complex geological situation is permissible but insists that the analysis of the model should be self-consistent in all detail.

The "model" referred to is that shown in Figures 5 and 6 and 19 and 20, where, in order to make a simplified calculation of the approximate maximum length of the overthrust block that could be pushed along a horizontal fault surface, rectangular blocks were shown resting upon a horizontal fault surface. The problem with which the authors were concerned was that of real overthrust faults in real rocks. In their phenomenological review (p. 120–122) they showed that these faults have been recognized most commonly in stratified rocks and consist typically of blocks of such rocks several kilometers thick and some tens of kilometers in length that have been slid over nearly horizontal fault surfaces.

Consequently, when it was desired to estimate approximately how long a block, several kilometers thick, could be pushed over a *horizontal fault surface*, a simple rectangular block of this thickness resting on a horizontal fault surface was shown. No further geological details were indicated in the drawings because these were not essential to the particular calculations that were to be made at this point in the discussion.

In the second case, where the buoyant effect of the contained water was introduced, again no stratification was shown in the drawings, but the real problem considered was still the geological one of water-filled sedimentary rocks, as was made quite clear by all the supporting evidence. For example, in Figure 25 (p. 151) a hypothetical sedimentary section was shown with a typical variation of the water pressure with depth. Then on pages 154–156 actual examples of such variation as measured in deep wells were cited.

In estimating the length of fault block for this case it was stated (p. 142–143)

". . . the rock is to have an interstitial fluid pressure p which varies as a function of the depth z."

Then later (p. 143), after λ had been substituted for p, and equation (95) had been derived for x_1, the maximum length of the block as a function of λ, it was added:

"In principle, λ may have any value between 0 and 1, and it may also vary as a function of depth. However, since slippage is favored along surfaces on which λ is a maximum, there is some justification for assuming that the value of λ at depths less than z_1 shall not exceed λ_1, the value at the base of the block."

For λ to vary as an arbitrary function of depth only it is essential that the rocks be horizontally stratified.

Thus, although no stratification was shown in the drawings the problem actually considered was that of a block of horizontally stratified sedimentary rocks in which λ varies as a function of depth and in which sliding is most likely to occur on a surface along which λ is a maximum. Since equation (95) could only be evaluated by assuming values for λ, it was pointed out that the minimum value for x_1 would be obtained if λ in the equation were assigned the maximum value λ_1, and this minimum length was computed. Actual faults with λ very probably less than λ_1 at depths less than z_1 should be longer than this minimum computed value given in Table 1.

The critic has interpreted these computational blocks literally. He has assumed that they represent the authors' "model" of the geological problem considered, and he criticizes the analysis for inconsistencies. Taking the blocks (Figs. 5 and 6, p. 127, Figs. 19 and 20, p. 143–144) literally as models, he assumes that they represent blocks of isotropic and homogeneous rock that are continuous with and have the same properties as the rock beneath. He then correctly points out that, if this were the actual case, the fault would not occur along a horizontal surface, but as an ordinary thrust

fault extending upward from the lower left-hand corner of the block.

Granting the critic's assumption, this would be true, but it is irrelevant for the problem actually investigated.

He comments that the unsupported vertical faces at the right-hand end of the blocks would collapse into some kind of an angle of repose. This also is true, but since the calculations are only approximate, and the strength properties of different actual fault blocks may vary through a range of several fold, the error of a few per cent resulting from computing a vertical-faced, rather than an inclined-faced, block is insignificant.

Birch also correctly points out that the authors neglected to show the downward-directed shear stress, increasing from zero at the top to a maximum at the base, on the left-hand ends of the blocks; and that, as a consequence, the normal components of stress σ_{zz} and σ_{zz} are not, as they assumed, principal stresses. A "rigorous calculation," according to Birch, for the "dry-rock" case of a block 5 km thick gives a length of 15.6 km instead of the authors' 18.4.

Details of the calculation are not given, but confirmation of Birch's results shows that he assumed the shear strength τ_0 to be effective simultaneously over the entire base of the block. Rejecting this assumption for reasons stated above, an even more rigorous calculation gives a length of 17.5 km, a reduction of only 5 per cent. Even this error tends to vanish when buoyancy is introduced, for in that case as λ_1 along the base increases, the shear stress τ approaches zero and σ_{zz} and σ_{zz} approach the principal stresses as limiting values.

Birch also argues that for the computed case of $\lambda = \lambda_1$, the overthrust block would be completely unstable. As has already been mentioned, this case was not assumed to exist geologically, but was considered only for computational purposes to obtain a lower limit for the lengths of blocks in which $\lambda < \lambda_1$. However, the instability argument still is not valid, for it is based upon the vanishing of τ_0 throughout the interior of the block whereas actually τ_0 would only vanish along the fault surface, as discussed heretofore.

A more serious error which the authors did not adequately take into account results from their neglect of an opposing end stress on the leading edge of the block. Once the block has detached itself and is being transported long distances it must override the topography in front, but in the initial stage it must somehow toe up at the leading edge, producing an opposing force to the movement. Neither did the authors explicitly point out in the original paper a fundamental difference between the stress patterns in a block that is being pushed from the rear and one that rests against a buttress at its lower end when tending to slide down an inclined plane. In the first case the longitudinal compressive stresses have a maximum value at the rear and diminish forward, whereas in the second case the same stresses have a maximum value at the front and diminish rearward.

This problem has been discussed in a paper entitled "Effect of the toe of a thrust plate," by David T. Griggs and C. B. Raleigh, which has been submitted for publication in the GSA *Bulletin*.

4. Concrete-Block and Beer-Can Experiments

We come finally to Birch's contention that both the concrete-block and the beer-can experiments are somehow invalid. He states that:

". . . The reduction of sliding friction by the presence of a pressurized liquid film which lightens or prevents contact between solid surfaces is unquestionable."

He then adds that this is not analogous to pore pressure, and that rocks and concrete have finite shear strength. Finally, he states unequivocally:

". . . There is no pore pressure in the beer can, nor in the concrete model."

He then attributes the results of the two experiments to the presence of a continuous liquid layer that completely separates the solid surfaces.

This raises two questions, the first concerning the mechanical interpretation of the two experiments, and the second their geological significance.

Concerning the mechanical interpretation, Birch is wrong when he asserts or implies that in each instance the reduction of friction is due to a film of liquid that completely separates the solid surfaces. In the beer-can experiment, which can easily be verified, the can slides down the glass plate making a continuous grinding noise of metal sliding on glass. Besides, the beer can, if entirely supported even momentarily by a liquid film, would be in a mechanically unstable position and would lean until contact between the metal and the glass was re-established and stability restored. Since

the experiment is performed twice, once with and once without the buoyant force of the air pressure inside, and since in each instance the materials are the same—wet beer can on wet glass—it follows that the only experimental variable capable of reducing the sliding friction is the force of contact between metal and glass. This is diminished as the air pressure is increased but is never reduced to zero. The reverse of this is shown dramatically by the air-brake effect that becomes operative the instant the can reaches the edge of the glass. Here the pressure is suddenly released, the brake is engaged, and the can is brought abruptly to a stop.

Thus, in this experiment, without altering the coefficient of friction between metal and glass, the angle θ of the slope down which the can will slide can be made to approach zero by decreasing the force of contact between the metal and the glass in accordance with the equation

$$\tan \theta = (1 - \lambda) \tan \phi , \qquad (10)$$

where ϕ is the angle of friction (about 20°), and $\lambda = F_p/W_n$ the ratio of the force of buoyancy to the normal component of the weight of the can.

In the concrete-block experiment the block was jointly supported by the rubber O-ring, the lower concrete block, and the water pressure, with two coefficients of friction, concrete-on-concrete (actually chert-on-chert because of the chert-pebble aggregate) and concrete-on-rubber. As was shown in the paper, the experiment had two phases, one in which the concrete blocks lost contact and the support was by the O-ring and water pressure, and the second where the concrete blocks retained contact. At no time did a solid-solid contact cease to exist. For the concrete-on-concrete the experimental results were in accord with equation (9) herein.

Here again, we have an experiment in sliding friction in which the stress across the solid-solid contact can be varied by variation of the pressure of the water entrapped in the pore space of the system; and, *without altering the coefficient of friction* as the experimental results plainly show, we may reduce the shearing stress required to slide the block by increasing the buoyant force exerted by the pressure of the water.

Despite their superficial differences, these two experiments are demonstrations of the same basic mechanics. Both show the drastic reduction in the shear stress that can be effected without altering the mechanical properties of the materials merely by increasing the fluid buoyancy. Their geological significance is obvious. The first shows the reduction of the force required to slide a block on a horizontal surface; the second shows the reduction of the angle of slope down which the body will slide under its own weight.

Birch is correct that the mechanics of these experiments differs from that of fault blocks by the absence of the term τ_0, but since, for the reasons discussed in Section 1 above, the term τ_0 in overthrust blocks can be neglected, its absence in the experiments is of no particular significance.

Conclusion

After having given serious consideration to Professor Birch's comments, the authors, while agreeing with a number of points he has made, still feel that:

(1) He is in error in his insistence upon the retention of the term τ_0 in the calculation of the maximum length of an overthrust fault.

(2) His interpretation of the buoyancy mechanism of reducing sliding friction as effecting a reduction of the coefficient of friction is wrong.

(3) His challenging of the validity of their approximate calculations of the length of overthrust-fault blocks because of lack of self-consistency with regard to unessential details is based on false premises.

When these aspects of Birch's arguments are disallowed, the changes that would be introduced into the authors' calculations become of second order. The authors still hold, therefore, to their original thesis, namely:

That, without postulating the existence of rocks with coefficients of friction or strength properties less than those observed for sedimentary rocks, the buoyant effect of the contained pore water at pressures such as those commonly observed in deep oil wells is sufficient to reduce the sliding friction of fault blocks to such an extent that the length of a block capable of being propelled by a thrust from the rear can be increased several fold, or the angle down which such a block can slide under its own weight can be reduced to but a few degrees.

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POSSIBLE TURBIDITY-CURRENT DEPOSIT IN NOVA SCOTIA

Abstract: The Meguma group, an apparently conformable sequence of slates and quartzites, extends over several thousand square miles with a thickness of 30,000 feet. Features indicating persistent current direction to the east, graded bedding, small scour and fill structures, convolute bedding, and large angular slate fragments in quartzite are described. Combined with their great lateral and vertical extent, these features fit well with an origin of deposition by turbidity currents.

A thick isoclinally folded sequence of interbedded quartzite and slate known as the Meguma group crops out over approximately 5000 square miles, or nearly one-third of the mainland of Nova Scotia. This sequence is bounded on the east and south by the Atlantic Ocean and on most of the west and north by a large granitic intrusive rock and unconformably overlying Mississippian sedimentary rocks. Because of the nature of the boundaries and the isoclinal folding, the total area of Meguma rocks is probably a few tens of thousands of square miles. The group has often been considered Precambrian, but there appears to be a conformable succession of strata in which some of the upper slates contain Ordovician fossils.

The Meguma has been divided into two formations. The lower, the Goldenville, consists of more than 16,000 feet of quartzite with interbedded slates. The upper, the Halifax, consists of approximately 14,500 feet of slate with some interbedded quartzite. Because the base is not exposed, the minimum thickness of the group has been considered as 30,000 feet (Malcolm, 1912).

The Goldenville quartzites, which range from gray to dark green, are arkosic in part, but many approach the composition of graywacke. Grits and conglomerate are found in limited amounts and appear at considerable depth in the formation. Slates in this formation range from gray to greenish black, and some are graphitic.

During the summer of 1956 several outcrops of the Goldenville formation along the Atlantic coast line northeast of Halifax between longitude 62° and 62½° West were examined.

In several outcrops on Moser River and on the eastern side of Liscomb Point some highly contorted strata are interbedded with layers showing no contortion. Certain of the distorted beds show less intense contortions toward the top. The structure of many layers is similar to the structure termed convolute bedding by Kuenen (1953, p. 1056).

Graded beds are found in outcrops along various rivers and are particularly well displayed along the coast at Beaver Point, Ball Gut, and Sober Island. These beds, each of which may be 6–24 inches in thickness, grade from coarse sandstone at the base through fine sandstone to shale, most of which has been converted to slate. At the base of most graded beds there are small scoured-out depressions, generally only a few inches across, extending into the underlying bed. These scours are filled with very coarse sandstone. Contacts between graded beds are sharp and distinct. In good exposures there may be 100 or more consecutive graded beds within approximately 100 feet stratigraphically.

In many outcrops, and particularly at Beaver Point, features that superficially appear to be cross-laminations are actually current ripple marks with their crests removed. In several of the graded beds the ripple marks occur in the coarser grades, but are not apparent from the fine sand size up to the top of the bed; this type of sequence can often be observed in several consecutive beds. Another persistent feature in the beds with ripple marks is the changing position of the crests upward through the bed. Wherever this advancing-crest feature is observed the advance is always in a consistent direction; where it can be observed in undisturbed outcrops it advances eastward. This agrees with the current direction indicated by the asymmetric ripple marks.

In many places angular fragments of slate occur in the quartzite. At Ball Gut a fragment in the lower part of a graded bed has a triangular cross section approximately 2 feet along each edge.

Several outcrops show wisps of an underlying layer curling up into the overlying layer. At Beaver Point there is a sequence of several of these wisps along the same interface. The current direction as indicated by ripple marks in

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the overlying bed suggests that these wisps correspond to the underthrust slump structure of Kuenen (1953, p. 1055–1956). Also at Beaver Point some of the outcrops contain what appear to be rolled-up mud balls that have a spiral cross section resembling jelly rolls.

Perhaps the most interesting features noted were U-shaped tubes approximately half an inch in diameter and several inches long in many of the quartzite beds. The two limbs of the U emerge at the surface of the bed. At the end of one limb there is a tentacular pattern radiating from a central hole. At the end of the other limb is an ellipsoidal mound with a hole in the center. There are hundreds of mounds and tentacular patterns on certain dip slopes, and they seem to have been caused by boring animals of some sort. Close observation indicates that the tentacular features are all curved in a manner that might be expected if they were swept by a current. The ellipsoidal mounds are rounded at one end and pointed at the other, similar to the cross section of an airplane wing. The current direction indicated by the swept tentacles agrees with that indicated by the airfoil shape of the mounds. Again, as in the case of the ripple marks, the current direction is toward the east.

Worm borings exist at Liscomb Point also, but these are not U-shaped. They appear as straight tubes and occur in slate rather than quartzite. On the south side of Liscomb Point is a bed containing branching algal fronds.

Near the crests and troughs of folds it is possible to follow the same sets of graded beds for several hundred feet in each of two perpendicular directions, but the lack of exposures and the nearly isoclinal folding make correlation over large distances difficult. The writer observed no pinching out or channeling.

Lateral persistence of beds containing current features all of which indicate the same current direction require a large-scale flow in which the direction of current is oriented. Deposits associated with the turbidity current produced by the 1929 Grand Banks earthquake have been well documented over thousands of square miles in the western Atlantic, and a uniform direction of flow seems well established (Heezen and Ewing, 1952; Ericson, Ewing, and Heezen, 1952; Heezen, Ericson, and Ewing, 1954). Large-scale deposits indicating uniform direction of flow are found in the eastern Pacific and also are attributed to turbidity currents (Menard, 1955). In addition to lateral extent of beds, origin of the above-described features must be accounted for. Some of these features are found in sediments resulting from laboratory-produced turbidity currents (Kuenen and Migliorini, 1950; Kuenen, 1953). None of the features would be inconsistent with such an origin. Thus, available evidence strongly suggests that the Meguma group was deposited, in large part, by turbidity currents.

There are very few petrographic or compositional data concerning these rocks. There has been no major publication about them since 1912 (Malcolm), and that was concerned primarily with overall structure and gold occurrences. Bell (1929, p. 24–25), in his study of the Horton-Windsor district of Nova Scotia, more than 100 miles west of the described area, briefly mentions rhythmically banded slates containing minute cross-bedding which he attributes to "rapid travelling of current ripples." He correlates these slates with the Meguma group. Such features agree well with those described in this paper.

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